Bulletin of the Geological Society of America

VOLUME 53 AUGUST 1, 1942 NUMBER 8

CONTENTS

PAGES

Endomorphism of the Idaho batholith. By Alfred L. Anderson.. 1099–1126

Pleistocene geology of western Cape Cod, Massachusetts. By Kirtley F. Mather, Richard P. Goldthwait, and Lincoln R. Thiesmeyer 1127–1174

Origin and structure of the Livingston igneous rocks, Montana. By Willard H. Parsons..... 1175–1186

Time relations in ocean sediments. By C. S. Piggot and Wm. D. Urry 1187–1210

Glaciation of Shickshock Mountains, Gaspe Peninsula. By R. F. Flint, Max Demorest, and A. L. Washburn..... 1211–1230

Mississippi delta—a study in isostasy. By Andrew C. Lawson.... 1231–1254

Subscription, $10 per year.

Publication Office: Florida Avenue and Eckington Place, Washington, D. C.

Communications for publication should be addressed to The Geological Society of America, Dr. H. R. Aldrich, Secretary, 419 West 117th Street, New York, N. Y.

NOTICE.—In accordance with the rules established by the Council, claims for non-receipt of the preceding number of the Bulletin must be sent to the Secretary of the Society within three months of the date of the receipt of this number in order to be filled gratis.

Entered as second-class matter in the Post-Office at Washington, D. C., under the Act of Congress of July 16, 1894.

Accepted for mailing at special rate of postage provided for in Section 1103, Act of October 3, 1917, authorized on July 8, 1918.

PRESS OF JUDD & DETWEILER, INC., WASHINGTON, D. C.

PAPERS IN PRESS FOR FORTHCOMING ISSUES

Pleistocene of Part of Northeastern Wisconsin. By F. T. Thwaites

Shiftings of the Mississippi River in Relation to Glaciation. By Frank Leverett

Paleozoic Paleogeography of Arizona. By Alexander Stoyanow

Flood Deposits of Arroyo Seco, Los Angeles County, California. By W. C. Krumbein

Wind-Polished Rocks in the Trans-Pecos Region, Texas and New Mexico. By Kirk Bryan and Claude C. Albritton, Jr.

"Pseudomigmatite" in the Piedmont of Maryland. By Randolph W. Chapman

National Research Council and Co-operation in Geological Research. By Walter H. Bucher

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 53, PP. 1099-1126, 3 PLS., 1 FIG. AUGUST 1, 1942

ENDOMORPHISM OF THE IDAHO BATHOLITH

BY ALFRED L. ANDERSON

CONTENTS

	Page
Abstract	1100
Introduction	1100
Divisions of the batholith	1102
Marginal facies	1103
Composition	1103
Rock description	1104
Mineral relations	1104
Interpretation of mineral relations	1107
Inner facies	1110
Composition	1110
Rock description	1110
Mineral relations	1111
Interpretation of mineral relations	1113
Aplite	1113
Distribution and structural relations	1113
Composition	1114
Rock description	1114
Mineral relations	1114
Interpretation of mineral relations	1115
Pegmatite	1115
Distribution and structural relations	1115
Composition	1116
Rock description	1116
Mineral relations	1117
Interpretation of mineral relations	1117
Endomorphism	1118
End-stage phenomena	1118
Endomorphism by post-consolidation solutions	1118
Endomorphic processes	1119
Relation of endomorphism and exomorphism	1119
Consolidation history	1119
Conclusions	1121
References cited	1121

ILLUSTRATIONS

Figure	Page
1. Sketch map showing distribution of Idaho batholith and other granitic rocks in the state	1101

Plate Facing page
1. Photomicrographs of mineral replacements in marginal facies..... 1124
2. Photomicrographs of mineral replacements in marginal and inner facies..... 1125
3. Photomicrographs of mineral replacements in inner facies and in pegmatite. 1126

ABSTRACT

Study of many parts of the Idaho batholith shows considerable endomorphism by post-consolidation solutions. The solutions, enriched particularly in silica but also in variable but generally appreciable quantities of potash and minor amounts of other elements, produced many changes in the solid rock. Biotite, orthoclase, microcline, muscovite, quartz, sphene, apatite, zircon, magnetite, garnet, allanite, epidote, zoisite, chlorite, and sericite replaced earlier minerals. As a result, the final rock is considerably more silicic and generally more alkalic.

Much of the dioritic marginal rock of the batholith has been changed to a quartz-rich diorite (tonalite) and granodiorite, and the less calcic rock of the interior from a quartz-bearing diorite and granodiorite to a somewhat calcic quartz monzonite with variations from granodiorite to granite. Dikes of associated aplite show similar end-stage changes, but most of the pegmatites have formed from somewhat later hydrothermal solutions by direct replacement of the aplitic and batholithic rock along fractures.

INTRODUCTION

For some time evidence has been accumulating that the rock of the Idaho batholith and adjacent similar masses (Fig. 1) has been considerably modified by post-consolidation emanations and that much of the intrusive masses is more silicic and alkalic than the original magma. Such modifications of the already consolidated rock were first noticed in granodiorite in the vicinity of Pend Oreille Lake in northern Idaho (Gillson, 1927, p. 1-20) and later over a much wider area in that region (Anderson, 1930a, p. 29-30; 1940, p. 17-21) and in the rock of the Cassia batholith in the extreme south of the State (Anderson, 1934a, p. 376-392). Mention of any such changes in the rock of the Idaho batholith, however, did not appear until 1934 (Shenon and Reed, p. 16), and no detailed descriptions until 1935 (Currier, p. 7-11) and 1940 (Anderson and Hammerand, 1940, p. 561-589), each report covering a small and widely separated part of the batholith. Anderson (1934b; 1939) also reported on other parts of the batholith but merely summarized the effects of the post-consolidation emanations.

The writer's interest in the consolidation of granitic rocks in Idaho began in the Clark Fork district near Pend Oreille Lake in 1927. The rocks had been considerably modified by post-consolidation emanations (Anderson, 1930a, p. 29-30), as in the near-by Pend Oreille district (Gillison, 1927, p. 1-20). Question arose as to whether other granitic masses, including the Idaho batholith, may not have been similarly affected. As field work in the State continued, the study was extended to the Idaho batholith and other granitic masses. Work on the batholith began

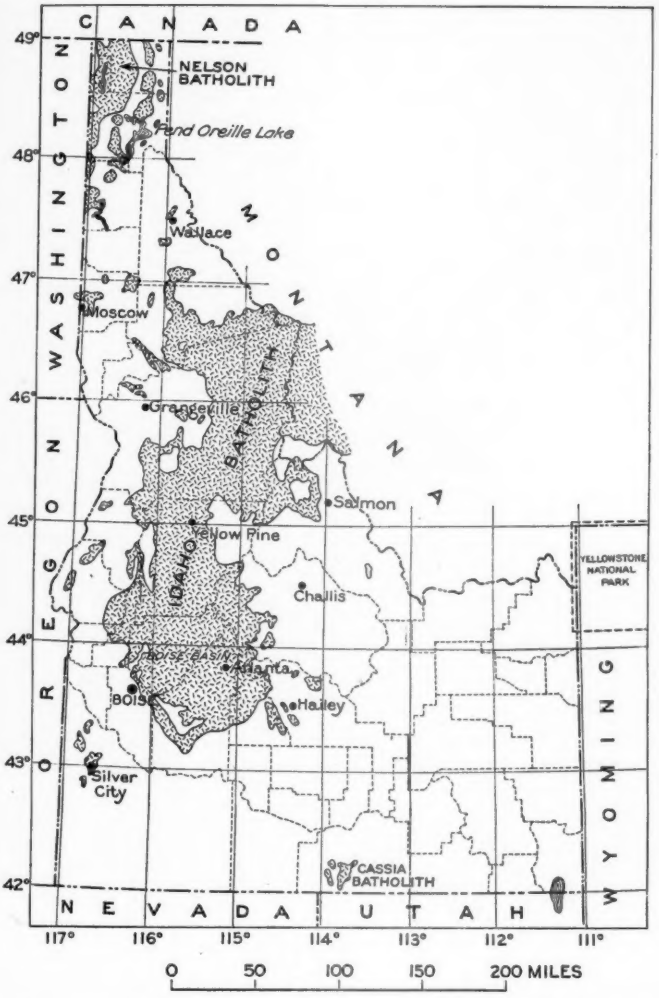


FIGURE 1.—*Sketch map showing distribution of Idaho batholith and other granitic rocks in the state*

in 1929 with examination of several thousand square miles of its northern part (Anderson, 1930b, p. 17-29). In 1930 the writer studied in detail the comparatively small Cassia batholith in southern Idaho, giving especial attention to effects of the igneous emanations on the invaded rocks (Anderson, 1934a, p. 376-392). Since 1931 much of the work on the batholith has been carried on in the south and southwest parts of the State, particularly in the Pearl-Horseshoe Bend (Anderson, 1934a, p. 1-36), Boise Basin (Anderson, 1942a), Atlanta (Anderson, 1939), and Rocky Bar (Anderson, 1942b) mining districts, but a small part of the batholith in the north part of the State (Anderson and Hammerand, 1940, p. 561-589) was also studied in detail.

Critical examination of many thin sections from all these districts as well as from other widely scattered localities in the Idaho batholith has made possible a fairly comprehensive analysis of the consolidation history of the entire batholith. Evidence from other parts indicates that the findings are applicable to the whole batholith.

In this paper only the endomorphism of the main granitic facies of the batholith and of associated aplite and pegmatite dikes is considered. The addition of igneous material to the invaded sedimentary rocks is not discussed as was done with the Cassia batholith (Anderson, 1934a, p. 376-392) and on a small part of the Idaho batholith in the north part of the State (Anderson and Hammerand, 1940, p. 561-589). Such studies in northern Idaho reveal that there is little material difference in the changes involved in progressive "granitization" of the invaded sedimentary rocks and in some of the changes induced in the granitic rock by endormorphic processes (Anderson, 1940, p. 17-20; Anderson and Hammerand, 1940, p. 561-589). However, the possible influence of incorporation of sedimentary material into at least the outer zones of the batholith cannot be wholly ignored in such speculations on the composition of the rock as are made herein.

The writer wishes to acknowledge his appreciation to G. F. Loughlin, C. S. Ross, and C. P. Ross for their assistance in the preparation of the manuscript and also the helpful suggestions made by J. C. Reed, P. J. Shenon, A. H. Koschmann, and others.

DIVISIONS OF THE BATHOLITH

Although the Idaho batholith, recognized as one of the major batholiths in North America, presents many interesting relations, this paper attempts to point out merely the broader structural and compositional features and to emphasize its natural division into two main facies differing from one another in age, structure, and composition. One facies is represented in a broad marginal and roof zone of more calcic rock consolidated under

considerable stress; the other is a less calcic inner zone which consolidated when stresses were less intense. In some localities more calcic rocks fringe the batholith, apparently intruded ahead of the main marginal mass, perhaps as injections of early little differentiated magma, but although the relations locally are complex, the presence of these subordinate masses does not rule out the picture of the batholith as consisting dominantly of two distinctive facies. This natural division into two main facies was first recognized as a major structural feature in the north part of the State (Anderson, 1930a, p. 17-29) and later in the southwest part where the distinction was based more particularly on differences in color and composition (Anderson and Rasor, 1933, p. 287-289). Since then others have noted the division; Ross, in the Casto quadrangle along a part of the east margin of the batholith (1934, p. 35-45), also later as a general feature (1936, p. 375-376); Currier, in the Yellow Pine district in the central part of the State (1935, p. 7-10); Shenon and Ross, in the Edwardsburg district also near the central part of the State (1936, p. 8-10); and Reed, in the Florence mining district near the west-central part of the batholith (1939, p. 7-8). In most places the two facies are transitional, but in some places the marginal rock is cut by the inner facies.

The marginal and inner facies differ appreciably in appearance and composition, and, in general, each may be readily identified. Locally the marginal facies is only a few hundred feet wide, but in most places it is thousands of feet or even miles across. The marginal facies makes up thousands of square miles of the exposed part of the batholith and equals roughly half the mass of the batholith. Because of the absence of systematic mapping, it is not feasible to show its distribution even approximately.

In places the marginal facies merges with sedimentary rocks that have undergone contact metamorphism, and it is difficult to tell how much of the marginal facies has a hybrid origin. Much of the rock discussed in this paper, however, appears to represent consolidated magma, and discussion of "granitized" borders is withheld.

MARGINAL FACIES

COMPOSITION

Much of the marginal facies is quartz diorite, but in broad areas it is granodiorite and locally quartz monzonite and granite. More calcic border rocks include peridotites, pyroxenites, gabbros, norites, and diorites, but since these escaped the general endomorphism of the main marginal facies, they are not considered further. The marginal facies is more calcic in the northwest part of the batholith, where over several thousand

square miles it is quartz diorite (Anderson, 1930b, p. 18). In the south and southwest part much of it is granodiorite (Anderson and Rasor, 1933, p. 287-294). In other parts of the batholith either or both types may prevail. Both quartz diorite and granodiorite are present in the Pearl-Horseshoe Bend region (Anderson, 1934b, p. 5-6). Granodiorite largely predominates in the Atlanta district (Anderson, 1939, p. 6-7), but quartz diorite also appears at Yellow Pine (Currier, 1935, p. 7-11), in the Casto quadrangle (Ross, 1934, p. 35-38), and in the Edwardsburg district (Shenon and Ross, 1936, p. 8-9). Quartz diorite predominates in the Elk City (Shenon and Reed, 1934, p. 12-13) and Florence (Reed, 1939, p. 7-8) areas. The variations from quartz diorite to granodiorite can, in general, be detected only microscopically and consequently are not mappable.

ROCK DESCRIPTION

Most of the marginal facies is moderately dark gray and somewhat gneissic. The light-colored grains commonly range from 4 to 7 millimeters in diameter, and the dark ones from 2 to 3 millimeters. Commonly 45 to 60 per cent consists of sodic andesine, averaging about $Ab_{65}An_{35}$. (In the inner facies the plagioclase is oligoclase.) The more calcic plagioclase in the marginal facies is generally distinctive. Quartz is everywhere conspicuous. In the northwest part of the batholith, it makes up 15 to 20 per cent, but in the southern and southwestern parts it is 25 to 30 per cent. Biotite and hornblende compose 10 to 15 per cent, rarely more. Biotite is commonly the more abundant, but hornblende is the principal dark constituent locally along the northwest margin. Small amounts of microcline are usually present, excepting in the northwest margin. Locally in the south the microcline increases up to 15 per cent, and even 30 per cent. Liberal sprinkling of grains of sphene and epidote is one of the distinctive features of the facies. Minor amounts of apatite, magnetite, zircon, allanite, zoisite, chlorite, muscovite, and sericite also are present.

MINERAL RELATIONS

Instead of the usual admixture of euhedral, subhedral, and anhedral forms, resulting from magmatic crystallization, many grains irregularly penetrate or impress one another in such a way that crystal faces are the exception but recognizable if the complete grain be restored by masking out the penetrating mineral.

Except for uneven penetration by biotite and quartz, hornblende tends to show crystal outline or outline modified by mutual interference during growth. Its outline is especially uneven along quartz contacts and appears as though indented by the quartz. If the quartz indentations were

filled with hornblende, however, the hornblende grain would appear in crystal outline.

The andesine also tends to have crystal outline, but because of its dominance over hornblende it has been much more affected by mutual crowding and is subhedral. However, contacts with microcline and quartz are extremely irregular; the microcline and quartz grains extend into and in places completely penetrate the andesine,—as they do the oligoclase in the inner facies (Pl. 2, fig. 3, 6; Pl. 3, fig. 1, 2, 3). Larger microcline grains enclose irregular remnants of andesine (Pl. 1, fig. 5). Although most andesine grains are clear and unzoned, some hold irregular ghostlike corroded inclusions of a more calcic, zoned plagioclase (Pl. 1, fig. 1). The calcic inclusions are rather widely scattered and appear to have been inherited, but from sources not determined. These zoned inclusions are ordinarily more highly sericitized than the host.

In general the biotite molds upon or lies between andesine and hornblende, but its shape does not conform altogether with the intergranular space. Some grains are unevenly penetrated by andesine and parts are retained in the andesine as small remnant oriented inclusions (Pl. 1, fig. 2). In some cases the penetrating andesine has fringelike borders as in Figure 2 of Plate 1, but more generally it impresses its own crystal outline against the biotite. On the other hand, some biotite grains penetrate the andesine, either maintaining essentially euhedral outlines (Pl. 1, fig. 3) or sending narrow protuberances into the andesine. Some grains penetrate one or more andesine crystals and some retain small oriented inclusions of the penetrated andesine. Thus the biotite has two habits, dependent on age, and is partly older and partly younger than andesine. Commonly the younger forms an added growth on the older (Anderson and Hammerand, 1940, p. 570). Borders against microcline and quartz are more regular; even crystal faces may show against the microcline. Against quartz, however, its borders, though smooth, are unevenly indented and grains may be cut entirely through (Pl. 1, fig. 4).

Microcline, if present in very scant amounts, generally forms roughly wedge-shaped grains among the andesine crystals; if more abundant, its grains become large and irregular. Deep embayments and irregular veinlets may then extend partly or entirely through the andesine crystals,—as they do in the oligoclase of the inner facies (Pl. 2, fig. 3). Reaction between microcline and andesine is suggested in many places by intervening zones in the andesine of differing extinction angles and composition. Wherever the microcline is more abundant and in still larger grains, it engulfs one or more andesine crystals as residual, corroded grains (Pl. 1, fig. 5), often as ghostlike shadows (Pl. 2, fig. 5). Crystal faces of hornblende and biotite determine the borders of the microcline. Euhedral

biotite or hornblende is not uncommon as inclusions in microcline. Its relations to the quartz are varied. In some specimens the microcline holds widely scattered, rounded, and irregularly corroded quartz grains,—similar to those appearing in the orthoclase of the inner facies—(Pl. 2, fig. 4) which appear to be remnants of anhedra originally wedged between plagioclase laths. Such quartz inclusions are exceptional, as most of the quartz is more or less lobate in and against the microcline (Pl. 3, fig. 2).

Quartz occurs also as small globular grains in andesine that has been corroded by microcline thus simulating myrmekitic intergrowths. Such intergrowths are quite prevalent throughout the facies. The largest part of the quartz occurs as irregular seams in feldspars and dark minerals or appears as large lobate grains which indent them. Except for the embayments into the plagioclase, hornblende and biotite grains, the quartz otherwise is largely interstitial. These relations are expressed in Figure 4 of Plate 1, which shows a seam of quartz cutting through a grain of biotite, also in Figure 6 of Plate 2 and in Figures 1, 2, and 3 of Plate 3. Except for Figure 4 of Plate 1, the photographs represent the inner facies, but the relations in the marginal rock are the same. Quartz seams of two ages are shown in Figure 3 of Plate 3, one cutting across and also embaying the other. Irregular remnants and lobes of one in the other are common. Thus two late stages of quartz deposition are indicated, which, added to the earlier two, make four in all, each with its own particular habit. The last two have like characteristics, except for age difference.

Much of the sphene occurs as large rhombic crystals cutting across or protruding into and through the hornblende, biotite, and andesine in a manner that does not accord with minerals crystallized in a pyrogenic sequence. The haphazard distribution of the sphene is partly controlled by grain contacts of other minerals (Pl. 1, fig. 6) and partly by mineral cleavages. Some of the sphene also occurs as large irregular grains, some of which have tongues projecting into adjacent grains, particularly into andesine (Pl. 2, fig. 1). Its habit therefore is largely penetrative, like much of the quartz. However, its relations to the quartz are indeterminate. It neither penetrates nor is penetrated by the quartz.

Grains and crystals of allanite, like sphene, penetrate the andesine and the biotite in a haphazard manner along grain contacts and mineral cleavages. Because it is less abundant than either the sphene or the epidote, its relations are not so clearly defined. Although usually occurring as independent crystals and grains, some is mantled or enclosed by epidote.

Scattered grains of epidote as large as the leading minerals are usually extremely irregular and thread through the biotite and hornblende crystals

and into the andesine (Pl. 2, fig. 2). In some places the epidote mantles grains of allanite, but most of it occurs as independent grains. Except for its irregular outlines, its relations resemble those of the sphene. Grains of zoisite appear more rarely scattered through the biotite and andesine.

Except for differences in size and amount, the grain relations of the scant accessories,—magnetite, apatite, and zircon—are like those of the sphene and allanite. The three minerals ordinarily occur in well-shaped, though small, crystals, aligned for the most part along borders of quartz, feldspar, and biotite grains, or scattered along fractures and cleavages in the feldspar or biotite.

Small grains of muscovite are widely scattered in the rock that contains large amounts of microcline. The grains ordinarily project into the andesine, less commonly into the microcline, and are closely associated with seams of quartz though contained in the bordering feldspar. Although the muscovite grains are relatively small, they are too large to be sericite and are easily distinguished from the minute grains of sericitic mica which are also common though not abundant. Much of the sericite occurs as small grains or aggregates of small grains in scattered patches in andesine. A little is found in thin seams which extend into and through the quartz grains.

Chlorite is not characteristic of the facies and is nowhere conspicuous in the fresh rock. Much of the chlorite represents partly altered biotite. It is subordinate and scarce as compared to epidote.

INTERPRETATION OF MINERAL RELATIONS

The mineral grains are intricately penetrated or irregularly penetrate neighboring grains so that their relations are not typical of orderly crystallization but resemble those of minerals formed by replacement found in alteration zones in wall rock bordering high-temperature ore deposits. Were most of the irregular penetrating grains masked by filling the embayments with the substance of the penetrated grains, euhedral, subhedral and interstitial auhedral grains would appear.

The hornblende does not have good crystal outline because it is penetrated by some of the grains of biotite, sphene, epidote, and quartz; and the andesine would be subhedral were its borders not so irregularly indented by biotite, microcline, and quartz. Some of the hornblende shows crystal faces against the andesine, and some andesine has crystal faces next to scattered wedge-shaped interstitial microcline and quartz grains. These interstitial grains are those that appear as inclusions in the larger grains of microcline. In general the biotite lies along and between the andesine, some of it penetrated by andesine and some of

it penetrating andesine or sending narrow protuberances into the andesine. Even though the sphene, allanite, magnetite, apatite, and zircon are euhedral in general they seem to be penetrating minerals.

On restoring the original crystal outlines the relations of the hornblende, andesine, and some of the biotite, potash feldspar, and quartz are those in ordinary consolidated magma.

Modification was probably orderly; minerals were added in rather definite sequence. The irregular protuberances from the biotite into and through the hornblende and andesine crystals suggest an added growth to grains originally between the hornblende and andesine crystals by replacement of the adjacent minerals. Late formation of biotite by replacement has been described by Gillson (1929, p. 77-86) in endomorphism in the Pioche district and by Anderson and Hammerand (1940) in a part of the Idaho batholith. Biotite is known to form in contact metamorphic zones in hydrothermal assemblages.

The microcline also indicates much replacement. Its extensive penetration of the andesine, accompanied in places by reaction rims of more silicic zones, and the presence in the microcline of numerous oriented remnants and shadowlike traces of andesine cannot be interpreted otherwise. On similar evidence Gillson (1927, p. 8) and the writer (1930a, p. 29-32) ascribed a replacement origin for the microcline in the granitoid rocks near Pend Oreille Lake; Currier (1935, p. 8-10) regarded the microcline in the Yellow Pine district as being added during the late stages of consolidation. In the Elk City region Shenon and Reed (1934, p. 16) pointed out such relations evidenced by reaction between microcline and plagioclase. Replacement of quartzite by microcline occurs in the contact zone of the Cassia batholith (1934a, p. 376-392) and the Idaho batholith in northern Idaho (Anderson and Hammerand, 1940, p. 575-579). The microcline shows no penetration nor corrosion of biotite but locally contains perfect crystals as inclusions, so it probably formed shortly after the late biotite and some of the pyrogenic quartz, retained as rounded and embayed inclusions.

The relations of much of the quartz also suggest its introduction by replacement. A minor amount has penetrated and replaced the andesine in myrmekiticlike blebs, but most of it occurs in large lobate grains, in seams, and in veinlets which penetrate the andesine, hornblende, and biotite, and in places the microcline. Its replacement relations seem to be identical to those of the quartz in the Sparta granite. Gilluly (1933, p. 65-81) emphasized the lobate habit of the quartz as one of the best criteria of replacement in the Sparta granite. Currier (1935, p. 8) regards the late microcline and quartz in the Yellow Pine district as contemporaneous, but the writer has observed that in some

sections the lobate quartz partly penetrates the margins of the microcline grains (Pl. 3, fig. 2). Seams and lobes of quartz are invaded by other seams and lobes of quartz, so that the late quartz is not all the same age.

The manner in which the sphene protrudes into and extends across grains of hornblende, biotite, and andesine, and its position in part controlled by fracturing and by cleavages, suggests replacement. Although irregular tongues may extend into and replace adjoining grains of andesine, the sphene commonly assumes its own crystal form during the replacement. Thus in its haphazard arrangement and other relations it is identical to the sphene in the granodiorite in the Pend Oreille region, which Gillson (1927, p. 5-8, 27-29) describes as a late introduction and replacement. The sphene in the granites in the Yellow Pine district is similarly regarded by Currier (1935, p. 8-11).

Much of the magnetite, apatite, and zircon is in small crystals concentrated along grain contacts and in fractures and cleavages of the earlier minerals, so that in occurrence they resembled the sphene. Gillson (1927, p. 16) pointed out these relations for the Pend Oreille district and Currier (1935, p. 8, 10) for the Yellow Pine district. They concluded that the minerals were added during a late replacement sequence, and the writer concurs.

Other minerals not already discussed likewise appear to be late replacements. The allanite has a distribution and habit not unlike that of the sphene and, as interpretated by Gillson (1927, p. 8, 27-29), has formed at the expense of the earlier minerals. The epidote occurs generally as large irregular grains with penetrating apophyses in and across cleavages of invaded minerals, and it, too, apparently is a late replacement similar to the late epidote described by Gillson (1927, p. 16-20). Its mantling of some of the allanite proves that the epidote is somewhat later. Chlorite occurs only as a minor alteration product of biotite in fresh rock, suggesting a late place in the sequence, as in the Pend Oreille district (Gillson, 1927, p. 16-20). The relations of the muscovite suggest replacement of the feldspars, possibly closely succeeding the formation of the microcline, but, as pointed out by Currier (1935, p. 8-10), as a phase of a still later stage in which sericite was formed. Narrow seams of sericite cutting the late quartz indicate its very late origin.

In addition to the sequence of minerals formed by crystallization another sequence is apparently superposed on the first by replacement. Thus to the succession of minerals of the early consolidated rock—the hornblende, biotite, andesine, in places a little potash feldspar, and generally scant amounts of quartz—are added the younger replacement sequence: (1) biotite, microcline, and quartz; (2) sphene, zircon, apatite,

and magnetite; (3) quartz, allanite, epidote, chlorite, muscovite, and sericite. The various stages may represent a continuous sequence. The order, particularly in the case of the minor accessories and the latest quartz, may have been somewhat different.

INNER FACIES

COMPOSITION

The inner facies has generally been regarded as comprising the prevailing rock of the batholith; it is a calcic quartz monzonite with local variations to granodiorite and granite which, with few exceptions, merge into one another. Such local variations are ordinarily discernible only in thin sections but are locally distinguished by abundant flesh-colored potash feldspar and much muscovite, which signifies change to granite.

This facies exhibits less regional variation in composition than the marginal facies, being uniformly alkalic and silicic throughout the batholith. Toward the north end this facies is typically quartz monzonitic with some variations to granite (Anderson, 1930b, p. 21). Farther south in the Elk City region Shenon and Reed observed quartz monzonite with almost equal granodiorite (1934, p. 16). Reed has found similar variations in the close-by Warren (1938, p. 7) and Florence (1939, p. 8) mining districts. Through other parts of the batholith the rock as at Edwardsburg (Shenon and Ross, 1936, p. 9), Yellow Pine (Currier, 1935, p. 8-10), Atlanta (Anderson, 1939, p. 7), Secesh (Capps, 1940, p. 7), and in the Casto quadrangle (Ross, 1934, p. 36) is mainly quartz monzonite, but granodiorite variations are locally almost as widespread as is the quartz monzonite. The facies is somewhat more calcic in the Boise Basin region. Much of it is granodiorite rather than calcic quartz monzonite (Anderson, 1942b).

ROCK DESCRIPTION

Most of the inner facies is much lighter-colored than the marginal having little or no evidence of gneissic structure which characterizes the marginal zone. Apparently the only exceptions so far observed are in the Yellow Pine (Currier, 1935, p. 8), Edwardsburg (Shenon and Ross, 1936, p. 9), and Casto (Ross, 1934, p. 36) areas where some slight banding has been pointed out. Much of the facies is typically light to medium gray, normally granitic but which may also be conspicuously porphyritic over broad areas, and in places somewhat aplitic. The light-colored grains commonly range from 4 to 7 millimeters in diameter, and the dark ones from 2 to 3 millimeters. The porphyritic rock differs from the normal only in being studded with Carlsbad-twinned potash feldspar phenocrysts as much as 30 millimeters long. In the

aplitic variations the grains range from 2 to 3 millimeters. Because the aplitic variety forms such a small proportion it is probably ordinarily overlooked. The only mention of it has been by Currier at Yellow Pine (1935, p. 11) and by the writer in Boise Basin (Anderson, 1942a).

Unlike the marginal the inner facies contains little or no hornblende, sphene, and epidote, commonly less than 5 per cent biotite, and has oligoclase (Ab_{75} to Ab_{80}) as its characteristic plagioclase. It also has considerably more potash feldspar—orthoclase and microcline—commonly muscovite, and generally larger amounts of quartz than the marginal rock. Its accessories include allanite, zircon, apatite, magnetite, and in places a little garnet. Other minerals include chlorite and sericitic mica.

Analyses of typical quartz monzonite from various parts of central Idaho (Ross, *in* Currier, 1935, p. 8) show the following range of composition in percentages: quartz, 25 to 30; potash feldspar, 16 to 35; plagioclase, 20 to 45; biotite, 1 to 10. The typical rock ordinarily contains in percentages: 20 to 30 quartz, about 20 of potash feldspar, a little more than 40 of plagioclase, and a little less than 5 of biotite. These proportions agree fairly well with those obtained by the writer in the Boise Basin and other parts of the batholith, except that in some places the quartz exceeds 30 per cent of the rock, and muscovite is as abundant as the biotite. Muscovite is not distributed throughout the rock of the facies but in various localities may be inconspicuous or absent altogether. More sericitic mica as well as considerably more chlorite occur in this facies.

MINERAL RELATIONS

Mineral relationships in the inner facies are much the same as those of the marginal quartz diorite and granodiorite. Among the essential minerals oligoclase alone tends to be euhedral or subhedral, but it has regular borders only against other oligoclase grains. Otherwise its borders show all stages of penetration by potash feldspar, biotite, and quartz. It also is commonly retained as irregular residual grains and oriented inclusions in the larger crystals of orthoclase and microcline. Most of the oligoclase grains are somewhat dusted with small grains of sericitic mica or hold patches of sericitic mica in aggregates of somewhat larger grains. The oligoclase crystals in places hold widely scattered ghostlike inclusions of a zoned, somewhat more calcic plagioclase.

Most of the other minerals penetrate older minerals. The biotite, as in the marginal rock, tends to be interstitial between the plagioclase grains, but in general its outlines are just as ragged as those of the biotite in the somewhat older quartz diorite and granodiorite either because it is pene-

trated irregularly by andesine or because it itself has irregular protuberances in the andesine (Pl. 2, fig. 3). The orthoclase and microcline also penetrate and corrode the oligoclase as the microcline corrodes the andesine in the marginal facies (Pl. 2, fig. 3). Myrmekitic quartz in the oligoclase, corroded by potash feldspar, is even more abundant and striking. Optical characters indicate that locally the oligoclase in contact with the potash feldspar changed in composition. Potash feldspar is considerably more plentiful in this facies than in the marginal zone; its grains are appreciably larger here, so that it contains many more plagioclase inclusions. Many of these inclusions are preserved as widely scattered but uniformly oriented small islandlike remnants or as shadowlike ghosts (Pl. 2, figs. 3, 5).

Where the potash feldspar grains are least numerous, they merely embay or cut through the oligoclase (Pl. 2, fig. 3), but where the grains are the size of phenocrysts they may hold remnants of several oligoclase grains as well as scattered, corroded grains of quartz, as typical "groundmass" inclusions (Pl. 2, figs. 4, 5). Megascopically the potash feldspar phenocrysts appear to be subhedral to euhedral and to be bounded by crystal faces; actually, as viewed in thin section, their borders are minutely irregular and have small protuberances projecting unevenly into adjoining plagioclase grains. In some places the potash feldspar is orthoclase, in other places it is wholly microcline; but usually orthoclase and microcline occur together, commonly forming parts of a single grain.

Only a small part of the quartz is included in and corroded by the potash feldspar. The bulk of it occurs as large grains which irregularly vein the rock and send lobate tongues into and through the oligoclase and biotite and to lesser extent into the potash feldspar (Pl. 2, fig. 6; Pl. 3, figs. 1, 2, 3). As in the outer facies this quartz contains irregular inclusions of earlier quartz, and seams and lobes of it are embayed and cut by other seams and lobes (Pl. 3, fig. 3).

Large muscovite grains occur in either the oligoclase or the potash feldspar, commonly at the side of quartz seams and lobes. Its outline is not nearly so irregular as the biotite grains. It apparently replaced the feldspars as euhedral grains.

Such minor accessories as magnetite, zircon, apatite, garnet, and allanite are mostly distributed along borders and cleavages of the feldspars and biotite and along margins of the quartz seams and lobes. Some of the chlorite forms entire or partial pseudomorphs after biotite. The sericitic mica occurs as scattered patches in the oligoclase and in lesser amounts in the potash feldspar. In places it appears as thin seams extending into and even across the quartz grains and seams.

INTERPRETATION OF MINERAL RELATIONS

Since the mineral relationships in the inner facies are practically identical to those in the marginal rock they are similarly interpreted. Were the oligoclase grains not so thoroughly embayed, corroded, and engulfed by the potash feldspar, they would possess partial crystal outline or one determined solely by crowding or mutual interference during crystal growth. Likewise, the biotite between and around the oligoclase grains suggests rather early crystallization, but its irregular penetrating borders argue also for a later growth. Wedge-shaped grains of orthoclase and quartz between some of the oligoclase crystals, or with the oligoclase as remnants within larger grains of corroding microcline, also suggest early crystallization. The following minerals therefore suggest normal, orderly crystallization from a consolidating magma: biotite, oligoclase, orthoclase, and quartz.

Another assemblage of minerals may be interpreted as replacements in the crystallized rock. The younger generation of minerals includes minor amounts of biotite, much orthoclase and microcline, two generations of quartz, and accessory apatite, zircon, and magnetite, in places allanite and garnet, and finally a little chlorite, and considerable muscovite and sericite. As the microcline commonly merges with orthoclase, probably the orthoclase formed earlier and at a higher temperature; both were deposited by replacement in a succession determined by temperature, the microcline perhaps by partial inversion of the orthoclase.

APLITE

DISTRIBUTION AND STRUCTURAL RELATIONS

Most of the aplite dikes are concentrated in relatively small areas in and adjacent to some of the mining districts within the batholith. The dikes are fairly numerous in the Gambrinus district in Boise Basin near Idaho City (Anderson, 1942a) and are exceptionally abundant in the Atlanta district (Anderson, 1939, p. 8). Numerous dikes are also reported near Yellow Pine (Currier, 1935, p. 7, 11-12), but in general aplites are scarce. Their distribution appears to be structurally controlled by certain zones of weakness developed during closing stages of consolidation of the batholith.

The aplitic bodies are ordinarily small,—usually several inches thick, exceptional ones in Boise Basin range up to 200 feet in thickness. The aplites in general are intrusions guided by fissures and ordinarily are tabular and have steep dips. Contacts between the dikes and the bordering walls are sharp.

COMPOSITION

As most of the aplites contain about equal potash feldspar and plagioclase they are quartz monzonite aplites. A few containing considerably more plagioclase than potash feldspar are granodiorite aplites. Whether any of the bodies contain a preponderance of potash feldspar and should be classed as granite aplites is a problem for future investigation. In some bodies grains of potash feldspar are larger and more abundant and express textural transition into pegmatites.

ROCK DESCRIPTION

The aplite is lighter-colored and considerably finer-grained than the prevailing rock of the batholith. In some of the larger dikes its average grains are about 2 millimeters in diameter, but in the smaller dikes grains average only about 1 millimeter or less. The unweathered rock is white, or pinkish, but the weathered somewhat resembles fine-grained brownish sandstone.

Most of the aplites apparently have about equal amounts of quartz, sodic oligoclase (Ab_{85} to Ab_{90}), and microcline and orthoclase, and less than 5 per cent of biotite, most of which is bleached or changed to muscovite. Some aplites contain sphene and also variable amounts of accessory garnet, zircon, apatite, allanite, and magnetite, and a little sericite, chlorite, and calcite. It differs essentially from the normal quartz monzonite and granodiorite in its finer grain, its occurrence, and in its more sodic plagioclase.

MINERAL RELATIONS

Although the texture is typically aplitic, the mineral relationships resemble those of the main rock of the batholith. Except for embayed margins, the plagioclase forms subhedral or euhedral crystals, and the biotite grains, if any protuberances were eliminated, would be largely subhedral. Most of the plagioclase grains are embayed by small lobes and seams of quartz, are penetrated and corroded by grains of potash feldspar, and are commonly penetrated by small slivers of biotite. The small amount of potash feldspar not invading other minerals forms small wedge-shaped grains between the plagioclase crystals. Some of the quartz does not embay or cut the plagioclase but forms wedge-shaped grains, many of which have later been engulfed in grains of microcline. Most of the accessory minerals are aligned along grain contacts and mineral cleavages. Some biotite is penetrated by or is replaced by chlorite. The biotite, chlorite, and feldspar are penetrated or replaced by muscovite and sericite. Some of the

sericite is scattered as granular aggregates in patches through the plagioclase. The calcite is in seams cutting the rock or in scattered grains.

INTERPRETATION OF MINERAL RELATIONS

Such small differences as may exist between the aplite and the normal rock are largely textural because of the finer grain and the presence of a somewhat smaller proportion of minerals that tend to penetrate other minerals. If the effects created by mineral penetration were eliminated, the rock would show euhedral and subhedral plagioclase, subhedral biotite, and anhedral orthoclase and quartz in the interstices. Thus the relations of minerals would be those in normally consolidated rock. These minerals, however, particularly the plagioclase, have been slightly penetrated by fringes of biotite, embayed and corroded by grains of potash feldspar, and cut and indented by lobes and seams of quartz. Also, the earlier grains have been penetrated along contacts and cleavages by small crystals of zircon, apatite, magnetite, and garnet, and some of them have been partly changed to chlorite, muscovite, sericite, and calcite. These changes must have been made by replacement of the minerals penetrated.

The aplites thus are the product of both crystallization and later replacements. After crystallization of plagioclase, biotite, and some potash feldspar and quartz another succession was added by replacement, a succession that included a little biotite, potash feldspar, quartz, and the various accessory and the so-called secondary minerals. Some biotite was added first, closely followed by potash feldspar and quartz. The different accessories were apparently deposited during the stage of quartz deposition, but, because the crystals were small and not in contact with each other, their precise order could not be determined. The chlorite, muscovite, sericite, and calcite sequence is more easily established and apparently marks the end of the replacements.

PEGMATITE

DISTRIBUTION AND STRUCTURAL RELATIONS

Pegmatitic seams and dikes are more widespread than the aplitic dikes but are more common at the roof and marginal zones of the batholith. Some of the pegmatites are related genetically to the marginal rock and in places are abundant in the outer facies. Others are related to the inner rock occurring at the outer margin of the inner facies and in the bordering marginal facies. The two series of pegmatites overlap in their distribution in the marginal zone and are not always easily segregated. Along the northwest margin of the batholith

the compositional contrast is clearly marked so that they may be separated, but in the southwestern part where no such contrast exists the two cannot be satisfactorily distinguished.

Where the pegmatites appear in the aplitic zones, as in the Atlanta district, they cut the aplite dikes or lie in and along them as though the fissures that guided the aplitic magma reopened at the time of pegmatite formation. Because of such reopening many of the aplites appear to grade into pegmatites. The pegmatites merge with or grade into the surrounding granitic and aplitic country rock.

COMPOSITION

Most of the pegmatites have about the same composition as the genetically related granitic rocks, and are designated as quartz monzonite pegmatite, granodiorite pegmatite, and quartz diorite pegmatite. Those related to the quartz diorite along the northwest margin of the batholith have been described as hornblende-quartz diorite pegmatite and biotite-quartz diorite pegmatite, and those associated with the inner quartz monzonite are quartz monzonite pegmatite and granite pegmatite (Anderson, 1930b, p. 19, 22). Most of the pegmatites associated with the marginal rock of the southwest range from granodiorite to quartz monzonite. Pegmatites related to the inner rock range from quartz monzonite to granite. In general the pegmatites tend to be alkalic, and several are considerably more alkalic than the parent facies and than other pegmatites in their own vicinity. Because the pegmatites around the northwest margin of the batholith have not been studied in as much detail as those in the southwest part they will not be discussed further; the following discussion deals with those examined in southwestern Idaho.

ROCK DESCRIPTION

Although graphic intergrowths of quartz and feldspar are not uncommon, most of the pegmatites resemble coarse-grained granite with the grains ranging from less than 1 inch to 6 inches in diameter. The dark minerals are invariably less abundant than in the related quartz monzonite and granodiorite, and as a result the rock is lighter. Some dikes are composed almost wholly of coarse crystals of pink or white feldspar and are uniformly white or pink, but in most dikes the rock is mottled because of many grains of smoky quartz and widely scattered crystals of black biotite.

The most abundant minerals are quartz, microcline, plagioclase, and small but variable amounts of biotite and muscovite. Most of the

dikes also contain scant amounts of apatite, zircon, magnetite, pyrite, locally garnet, and chlorite and sericite. In pegmatites associated with the marginal facies, the plagioclase is andesine; in other pegmatites, generally oligoclase, but in some it is albite. The proportions of the essential minerals are variable. Quartz is relatively abundant in most of the dikes; except in the pegmatites related to the marginal facies, the microcline is usually more plentiful than the oligoclase. Many of the highly feldspathic dikes are composed almost entirely of microcline. Widely scattered crystals of biotite and muscovite, ranging up to 3 inches in diameter, are more common in the more siliceous pegmatites. Other minerals, except locally small crystals of garnet, are as a rule very inconspicuous.

MINERAL RELATIONS

Most of the plagioclase and some of the grains of quartz are completely enclosed in the microcline (Pl. 3, fig. 4),—some of the enclosed grains altered to shadowlike inclusions. These inclusions commonly match the plagioclase and quartz grains in the adjacent granite and aplite and actually merge with them at the contact, the grains of microcline penetrating and isolating remnants of the invaded rock. The plagioclase grains are also penetrated by crystals of biotite which may even project into the wall rock. The biotite, however, is not penetrated by the microcline. Albite forms perthitic spindles in the microcline or crystals which project into the microcline grains. Although some of the quartz occurs as residual grains in the microcline, most of it is in grains and seams that penetrate the biotite, plagioclase, and microcline. It is lobate like much of the quartz in the marginal and inner facies. Some of the quartz cuts the large microcline grains in irregular veinlets that locally simulate graphic intergrowths. The muscovite and most of the other minerals lie almost wholly within the feldspar grains.

INTERPRETATION OF MINERAL RELATIONS

With the exception of the inclusions of plagioclase and quartz that obviously were derived from the older rocks, the writer believes all the minerals were formed by replacement. The biotite that projects through the residual masses of included wall rock, the microcline that retains numerous residual inclusions of the country rock, the albite that penetrates the microcline, and the lobate quartz that cuts and embays them all, permit no other interpretation. Deposition of the quartz apparently persisted until after the muscovite and all the accessory minerals had been added.

ENDOMORPHISM

END-STAGE PHENOMENA

Mineral relations of the marginal and inner facies and of the associated aplite and pegmatite dikes show that the present rock has resulted partly from replacement. Each facies and associated aplites, except the pegmatites, have had much the same minerals added. The pegmatites differ because the replacements have not been superposed upon crystallized minerals of a pegmatitic magma but have acted upon the minerals of the inner and marginal facies and of the aplite.

Such changes have been cited by Gillson (1927, p. 5-8, 16-30), Currier (1935, p. 8-11), and Anderson and Hammerand (1940, p. 585-586) as end-stage replacements, taking place during the closing stages of consolidation. Considerable but variable amounts of potash and silica and minor amounts of other materials have been contributed, which must have come from the batholith as emanations from the deep magma.

ENDOMORPHISM BY POST-CONSOLIDATION SOLUTIONS

Gillson (1927, p. 16-30) and Currier (1935, p. 8-9), who have described the changes in parts of the batholith regard some of the changes as deuteric forming by reaction of the crystals with residual liquid. Gillson then has described subsequent changes as pneumatolytic and hydrothermal, and Currier, as wholly hydrothermal. It appears to the writer that the late assemblage of biotite, potash feldspar, quartz, sphene, zircon, apatite, magnetite, allanite, epidote, garnet, chlorite, muscovite, and sericite is typically hydrothermal and accords with a gradual decline in temperature. Possibly the reaction rims on some of the plagioclase and the myrmekitic blebs of quartz in the plagioclase grains may have been produced by the reaction of crystals with residual liquid (Currier, 1935, p. 8-9); however, since these deuteric effects are so closely associated with replacement, they may have been produced during the succeeding post-consolidation stage. In view of the considerable amounts of material added to the rock, the changes must have taken place primarily under the influence of hydrothermal solutions. These solutions were probably residual liquids in the magma at a somewhat greater depth, perhaps at much greater depth in the case of the solutions which produced the pegmatites. Because of the difficulty of distinguishing between alteration of deuteric origin and replacement during an immediately succeeding high-temperature hydrothermal stage, and since any relation that might be construed as deuteric constitutes such a minor part of the magmatic after-effects, no distinction between the deuteric and the hydrothermal is made in this paper. The stages of alteration

may have been represented by a continuous sequence conforming to a gradual change in temperature and chemical character of the hyperfusible constituents which became concentrated during the later stages of consolidation and then emanated from depth while the cooling, consolidation, and fracturing of the batholith proceeded downward.

ENDOMORPHIC PROCESSES

Such changes as have taken place in the consolidated, or largely consolidated, rocks might be described as the product of end-stage metamorphism or post-consolidation metamorphism. However, the term endomorphism accords more with common usage. The main process of endomorphism has been silicification as it has been in the Sparta granite (Gilluly, 1933, p. 65-81). Large amounts of silica have been added to the rock of the Idaho batholith, and there has also been a variable enrichment in potash. The addition of potash feldspar has caused compositional variations within both facies and in the aplite and pegmatite.

RELATION OF ENDOMORPHISM AND EXOMORPHISM

Where emanations have invaded country rock, changes quite analogous to those within the intrusive rock have been produced. If the invaded rocks were quartzitic they have become rather extensively granitized. Such contact effects have been described around the margin of the Cassia batholith (Anderson, 1934a, p. 376-392) and were a chief factor in the development of the batholith in the Selkirk Mountains in northern Idaho where emanations have changed the Belt sedimentary rocks almost entirely into a distinctly banded gneiss ranging from granite to quartz diorite (Anderson, 1940, p. 17-20). Similar contact phenomena also have been described for the Idaho batholith in north Idaho (Anderson and Hammerand, 1940, p. 561-589). The problems of endomorphism are probably closely related to those of the contact metamorphism and granitization.

CONSOLIDATION HISTORY

The consolidation history of the Idaho batholith has been complicated by the two-stage emplacement. In spite of the solidification of the earlier marginal rock under somewhat greater stress than the later inner facies and despite the fact that a time interval of unknown duration permitted an appreciable change in the composition of the inner mass before its emplacement and consolidation, the endomorphism of the rock of each facies has been remarkably similar. Evidence indicates that the original marginal facies was a diorite with only minor amounts of quartz but locally with hornblende or biotite, or both, and

with andesine. As a result of end-stage contributions, some biotite was added, as were small but variable amounts of potash feldspar, much quartz, generally exceptionally large amounts of sphene, and such minor minerals as apatite, zircon, magnetite, allanite, epidote, chlorite, locally muscovite, and sericite. The most important contribution was silica, and the rock was in general thoroughly silicified. Where little or no potash feldspar was added, as along the northwest margin of the batholith, a quartz-rich diorite (tonalite) was produced, but where considerable amounts of potash feldspar were added, as along the south and southwest margin, granodiorite formed.

The consolidation history of the inner facies differs but slightly in detail from that of the marginal. Its magma was apparently somewhat less calcic than that of the marginal rock and yielded an original rock ranging apparently from diorite or quartz-bearing diorite to granodiorite with oligoclase, rather than andesine. Post-consolidation emanations added considerable silica and potash, increased slightly the amount of biotite, and introduced the materials for the formation of the accessories. More potash feldspar (orthoclase and microcline) was added than in the marginal rock, thus changing much of the original rock to quartz monzonite. Where less potash feldspar was added, the conversion was to granodiorite. Locally enough potash was added to form granite, in places a muscovite granite. Over wide areas the added feldspar formed replacement phenocrysts that make the rock conspicuously porphyritic. As in the marginal rock silicification has been the dominant process in the endomorphism, but the process by which potash feldspar was added also has been important.

Consolidation of the aplite dikes has followed a course similar to that of the main granitic facies. Progressive cooling of the batholith inward and downward apparently was accompanied by shrinking and cracking of the solidified roof zone. When avenues were opened to greater depth by deeper-seated fissuring, the magma, which had become less calcic, perhaps because of longer-continued differentiation, was injected upward into the fractures near the top, where it consolidated as aplite. Then the aplite was considerably endomorphosed by associated emanations and was made somewhat more silicic and alkalic.

Pegmatites, on the other hand, apparently did not originate directly from consolidating magma as did the aplites and granitic facies but appear to be wholly hydrothermal and associated with replacement of the earlier consolidated granitic and aplitic rock, by the action of silica and potash-rich fluids from deeper sources than those that endomorphosed the aplites.

CONCLUSIONS

That the composition of intrusive igneous rocks may be considerably modified by the action of end-stage emanations is rather generally recognized, but the extent to which such action might affect huge batholithic bodies has probably not been fully appreciated. The endomorphism that accompanied the consolidation of the Idaho batholith may be more than locally significant and may be a feature associated prominently with batholiths in other localities.

REFERENCES CITED

Anderson, A. L. (1930a) *Geology and ore deposits of the Clark Fork district, Idaho,* Idaho Bur. Mines Geol., Bull. no. 12, 132 pages.

—— (1930b) *Geology and mineral resources of the region about Orofino, Idaho,* Idaho Bur. Mines Geol., Pamph. no. 34, 63 pages (mimeographed).

—— (1934a) *Contact phenomena associated with the consolidation of the Cassia batholith, Idaho,* Jour. Geol., vol. 42, p. 376-392.

—— (1934b) *Geology of the Pearl-Horseshoe Bend gold belt, Idaho,* Idaho Bur. Mines Geol., Pamph. no. 41, 36 pages (mimeographed).

—— (1939) *Geology and ore deposits of the Atlanta district, Elmore County, Idaho,* Idaho Bur. Mines Geol., Pamph. no. 49, 71 pages (mimeographed).

—— (1940) *Geology and ore deposits of Kootenai County, Idaho,* Idaho Bur. Mines Geol., Pamph. no. 53, 67 pages (mimeographed).

—— (1942a) *Geology and ore deposits of Boise Basin, Idaho,* U. S. Geol. Survey, Bull. (in press).

—— (1942b) *Geology and ore deposits of the Rocky Bar district, Elmore County, Idaho,* Idaho Bur. Mines Geol., Pamph. (in press).

—— and Hammerand, V. (1940) *Contact and endomorphic phenomena associated with a part of the Idaho batholith,* Jour. Geol., vol. 48, p. 561-589.

—— and Rasor, A. C. (1933) *Composition of a part of the Idaho batholith in Boise County, Idaho,* Am. Jour. Sci., 5th ser., vol. 27, p. 287-294.

Capps, S. R. (1940) *Gold placers of the Secesh Basin, Idaho County, Idaho,* Idaho Bur. Mines Geol., Pamph. no. 52, 43 pages (mimeographed).

Currier, L. W. (1935) *A preliminary report on the geology and ore deposits of the eastern part of the Yellow Pine district, Idaho,* Idaho Bur. Mines Geol., Pamph. no. 43, 27 pages (mimeographed).

Gillson, J. L. (1927) *Granodiorites of the Pend Oreille district of northern Idaho,* Jour. Geol., vol. 35, p. 1-20.

—— (1929) *Petrography of the Pioche district, Lincoln County, Nevada,* U. S. Geol. Survey, Prof. Paper 158-D, p. 77-86.

Gilluly, James (1933) *Replacement origin of the albite granite near Sparta, Oregon,* U. S. Geol. Survey, Prof. Paper 175-C, p. 65-81.

Reed, J. C. (1938) *Geology and ore deposits of the Warren mining district, Idaho County, Idaho,* Idaho Bur. Mines Geol., Pamph. no. 45, 65 pages (mimeographed).

—— (1939) *Geology and ore deposits of the Florence mining district, Idaho County, Idaho,* Idaho Bur. Mines Geol., Pamph. no. 46, 44 pages (mimeographed).

Ross, C. P. (1934) *Geology and ore deposits of the Casto quadrangle, Idaho,* U. S. Geol. Survey, Bull. 854, 135 pages.

——— (1936) *Some features of the Idaho batholith,* XVI Internat. Geol. Cong. Rept., Washington, p. 369-385.

Shenon, P. J., and Reed, J. C. (1934) *Geology and ore deposits of the Elk City, Orogrande, Buffalo Hump, and Ten Mile districts, Idaho County, Idaho,* U. S. Geol. Survey, Circular no. 9, 89 pages (mimeographed).

——— and Ross, C. P. (1936) *Geology and ore deposits near Edwardsburg and Thunder Mountain, Idaho,* Idaho Bur. Mines Geol., Pamph. no. 44, 45 pages (mimeographed).

Department of Geology, Cornell University, Ithaca, New York.
Manuscript received by the Secretary of the Society, October 31, 1941.
Presented before Cordilleran Section, June 17, 1940.
Published with permission of the Director, U. S. Geological Survey.

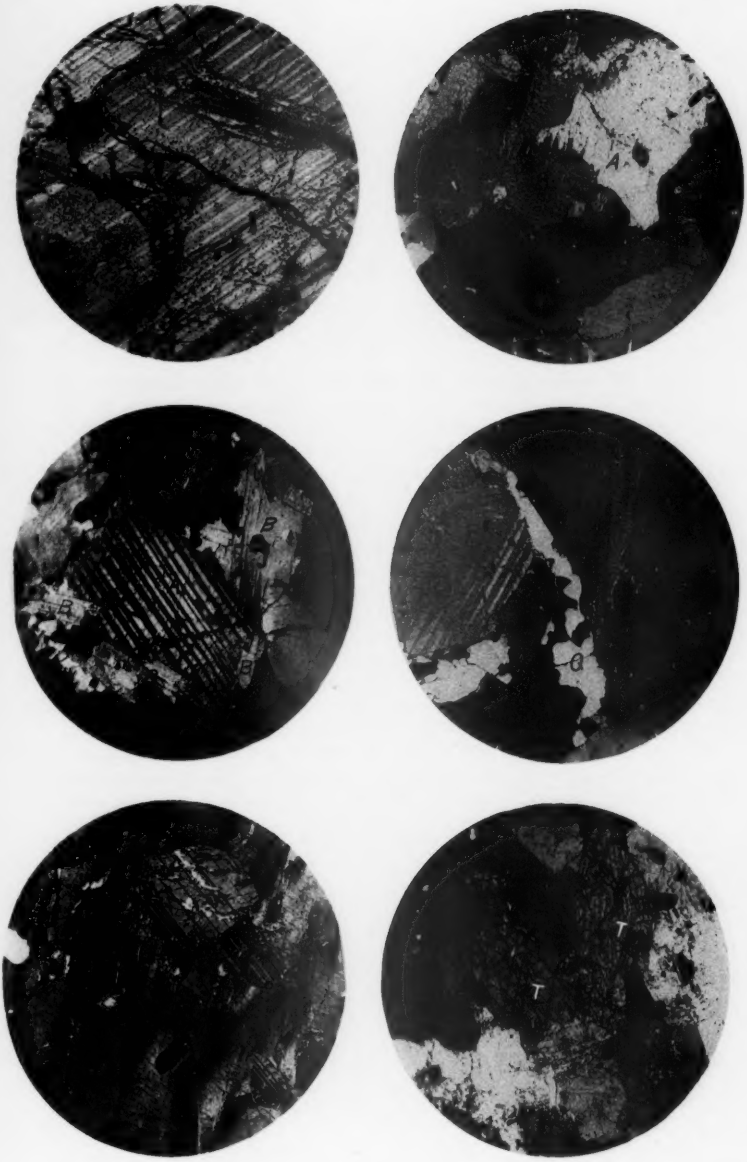
EXPLANATION OF PLATES

PLATE 1

PHOTOMICROGRAPHS OF MINERAL REPLACEMENTS IN MARGINAL FACIES

Figure

1. Andesine crystal containing irregular remnant of an earlier, more calcic andesine Crossed nicols, × 28.
2. Biotite (B) showing irregular penetration by andesine (A) Oriented islands of the biotite remain as inclusions. This biotite is interpreted as early and as partly replaced by the andesine. Generally, contact between invading andesine and biotite is more regular. Biotite may have been inherited from "granitized" quartzite. Crossed nicols, × 28.
3. Andesine crystal (A) irregularly penetrated at its margin by crystals of biotite (B). Biotite crystal in northeast quadrant also penetrates and partly replaces a second andesine grain. The biotite is interpreted as a late addition to the rock during a post-consolidation stage of alteration. Crossed nicols, × 37.
4. Biotite (B) between andesine grains (A) cut and partly replaced by a vein of late post-consolidation quartz. Crossed nicols, × 28.
5. Microcline which contains remnants of partly replaced andesine grains. Crossed nicols, × 28.
6. Sphene crystals grouped between and along feldspar grains. The sphene assumes its own crystal form in replacing other minerals. Replacement interpreted as taking place during an early stage of post-consolidation alteration Crossed nicols, × 37.



PHOTOMICROGRAPHS OF MINERAL REPLACEMENTS IN MARGINAL FACIES

PHOTOMICROGRAPHS OF MINERAL REPLACEMENTS IN MARGINAL AND INNER FACIES

PLATE 2

PHOTOMICROGRAPHS OF MINERAL REPLACEMENTS IN MARGINAL AND INNER FACIES

Figure

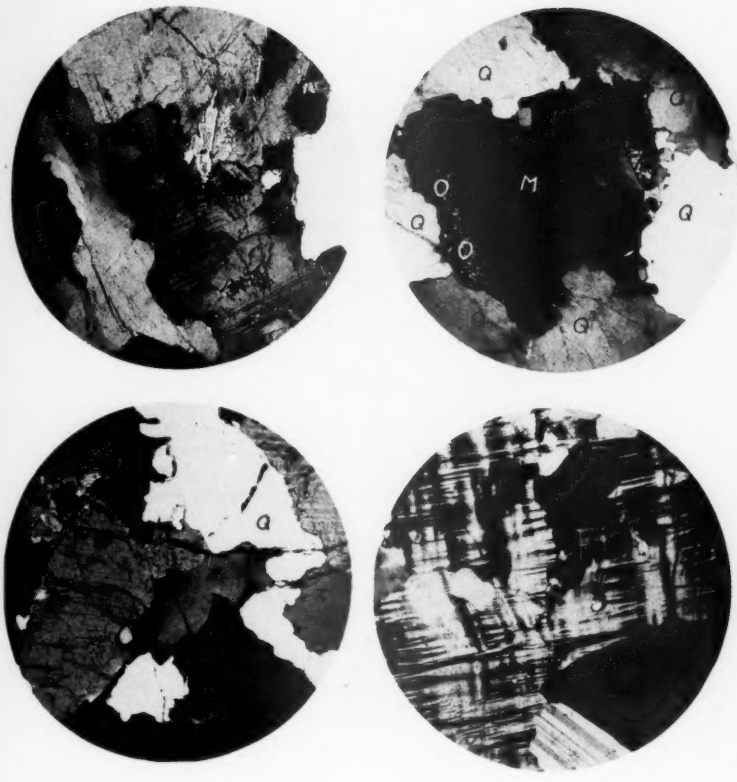
1. Andesine crystal (A) partly included in a large crystal of sphene (T). Apophysis of the sphene also extends outward into another grain of andesine (right). Sphene interpreted as added to the rock during an immediately succeeding stage of post-consolidation alteration. Crossed nicols, × 28.
2. Large grain of epidote (E) with apophyses extending into and replacing a crystal of andesine (A). Affords an example of post-consolidation replacement in marginal facies. Crossed nicols, × 28.
3. Biotite (B) projecting into and partly replacing oligoclase crystal (0) and in turn partly embayed by microcline. Vein of microcline (M) also cuts entirely through the oligoclase crystal, isolating a part of it from the main crystal. Example of replacement phenomena in the inner facies. Crossed nicols, × 37.
4. Large grain of orthoclase with remnant inclusions of earlier oligoclase (0) and quartz (Q). Orthoclase is interpreted as an addition to the rock of inner facies during period of end-stage alteration. Crossed nicols, × 37.
5. Large microcline crystal with remnant quartz grains (light) and shadowlike inclusions of original plagioclase feldspar. Microcline is interpreted in part as a replacement of the inner facies. Crossed nicols, × 37.
6. Shows the way the late-stage quartz (Q) embays the oligoclase (0). Replacing lobes may penetrate far into the interior of the oligoclase. Crossed nicols, × 28.

PLATE 3

PHOTOMICROGRAPHS OF MINERAL REPLACEMENTS IN INNER FACIES AND IN PEGMATITE

Figure

1. Seam of lobate quartz (Q) penetrating in and between crystals of oligoclase. Uneven borders of quartz grains suggest replacement rather than interstitial filling. Typical of much of the late quartz of the batholith. Crossed nicols, × 28.
2. Microcline grain (M) penetrated unevenly on three sides by lobate quartz (Q) and protected on fourth side by similarly penetrated grain of oligoclase (O). Crossed nicols, × 37.
3. Seam of early quartz (Z) cut by seam of somewhat younger quartz (Q). Both penetrate and replace oligoclase grains. The younger quartz (Q) also embays and replaces the older quartz (Z). Two late stages of quartz deposition is suggested. Crossed nicols, × 28.
4. Microcline grain in pegmatite containing remnant inclusions of quartz and oligoclase inherited from the quartz monzonite. Microcline is interpreted as replacing the wall rock. Crossed nicols, × 28.



PHOTOMICROGRAPHS OF MINERAL REPLACEMENTS IN INNER FACIES AND IN PEGMATITE

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 53, PP. 1127-1174, 7 PLS., 4 FIGS. AUGUST 1, 1942

PLEISTOCENE GEOLOGY OF WESTERN CAPE COD, MASSACHUSETTS

BY KIRTLEY F. MATHER, RICHARD P. GOLDTHWAIT, AND LINCOLN R. THIESMEYER

CONTENTS

	Page
Abstract	1128
Introduction	1128
Work on which this report is based	1128
Acknowledgments	1129
Earlier work on Cape Cod	1129
Part I. Glacial moraines in the vicinity of Buzzards Bay	1131
Buzzards Bay moraine	1131
Sandwich moraine	1143
Relations between Sandwich and Buzzards Bay moraines and Mashpee pitted plain	1146
Overlap of Sandwich moraine upon Buzzards Bay moraine	1147
Scorton moraine	1148
Ellisville moraine	1150
Summary	1150
Part II. Mashpee pitted plain	1151
Surface features and composition	1151
Glacial origin	1153
Relations to ice fronts	1157
Till clumps	1159
Subaerial furrows	1160
Persistence of buried ice	1161
Drowned shore line	1162
Part III. Late glacial ventifacts	1163
Occurrence and distribution	1163
Rock types represented	1165
Sizes and shapes	1166
Evidence of sandblasting	1167
Abrasion during subsequent transportation	1170
Wind-cut surfaces transecting weathered rims	1170
General Statement	1170
Date of formation of weathered rims	1171
Significance of ventifacts in dating deposits	1171
Climatic implications	1172
Summary	1173
References cited	1173

ILLUSTRATIONS

Figure	Page
1. Outline map of southeastern Massachusetts	1130
2. Map showing glacial geology of western Cape Cod	1132

Figure — Page

3. Sketch map of western Cape Cod 1152
4. Profiles across Mashpee pitted plain 1154

Plate — Facing page

1. Glacio-lacustrine deposits in Brick Kiln Road pit 1138
2. Materials composing Mashpee pitted plain 1139
3. Ventifacts photographed in the field 1164
4. Pyramidal ventifacts and wind polish 1165
5. Influence of rock texture on formation of ventifacts 1168
6. Grooved and etched ventifacts 1169
7. Ventifacts of various shapes 1170

ABSTRACT

During recession of continental ice of the Wisconsin stage a complex series of moraines and pitted plains was deposited on Cape Cod and adjacent portions of southeastern Massachusetts. Most notable are the Buzzards Bay moraine, extending in a northeast-southwest direction along the southeast shore of Buzzards Bay, and the Sandwich moraine, trending east-west and roughly parallel to the south shore of Cape Cod Bay. These were deposited during minor readvances of the lobate margin of the ice. The lithologic character of the transported stones reveals the direction from which the ice advanced and is distinctive for the two major lobes of ice. Along the shore north of the Sandwich moraine there are discontinuous bodies of till resting upon and surrounded by glacio-fluvial deposits. These constitute the Scorton moraine, deposited during the wastage of essentially stagnant ice, after retreat of the ice-front from the position marked by the Sandwich moraine. The Ellisville moraine, north of Buzzards Bay, represents a third type of recessional moraine, resulting from a temporary halt in the retreat of the margin of an active ice-lobe without readvance.

The Mashpee pitted plain occupies the angle between the Buzzards Bay and Sandwich moraines. It consists largely of sand and gravel spread by melt water across a land surface strewn with many large blocks of stagnant ice, left stranded beyond the continuous, lobate margin of the ice sheet, at a time when that margin was an appreciable distance north and northwest of the line now marked by those moraines and prior to the readvance of the ice responsible for their emplacement. An extraordinary number of large and deep kettle holes dot the surface of the gently sloping fan-shaped plain. Adjacent to some of them, clumps of till are embedded in the glacio-fluvial deposits. Many of the buried blocks of ice responsible for the kettle holes did not melt away until after the surface of the plain had been extensively furrowed by subaerial erosion and its western and northern margins had been overridden by active ice readvancing to form the bordering moraines.

Thousands of ventifacts are embedded in undisturbed glacial and glacio-fluvial deposits and in disturbed surficial zones throughout the region. All types of wind abrasion from incipient polishing and faceting to pronounced etching, deep fluting, or pitting are displayed. These stones were sculptured by sandblasting in a periglacial climate, both prior to the deposition of the till and during the construction of the glacio-fluvial deposits. Many of the wind-scoured surfaces were modified to greater or lesser extent by abrasion during glacial or glacio-fluvial transport. Thin weathered rims on some of the ventifacts apparently survived such transport but were truncated by wind-cut facets.

INTRODUCTION

WORK ON WHICH THIS REPORT IS BASED

Field investigations in the western part of Cape Cod were begun in June 1939 under the auspices of the United States Geological Survey as a part of a co-operative project for geological research in Massachu-

setts, financed jointly by the Commonwealth and the Federal Government. Recently completed topographical maps with a scale of 2 inches to the mile were used as the base for the detailed geologic mapping which was extended over most of the area covered by the Sagamore, Sandwich, Pocasset, Falmouth, and Woods Hole quadrangles and the eastern part of the Onset quadrangle during the summer of 1939.

ACKNOWLEDGMENTS

During the progress of the field work and preparation of this report, the writers were greatly aided by numerous conferences in the field and office with L. W. Currier, geologist in charge of the co-operative project, Robert W. Sayles, and Kirk Bryan. Arthur S. Knox studied many localities and is responsible for the identification of the animal and plant remains which give a clue to the fauna and flora of the region during late Pleistocene and Recent time. Edward A. Schmitz prepared the maps and diagrams for publication. For the photographs of ventifacts taken in the laboratory the writers are indebted to Edward S. Wood, Jr., of the Institute of Geographical Exploration at Harvard University.

EARLIER WORK ON CAPE COD

The glacial, glacio-fluvial, and glacio-lacustrine deposits comprising nearly all the land mass known as Cape Cod have long been of great interest to geologists and have been the subject of many reports published during the last hundred years. The abundant erratics strewn over the surface of the Cape and the adjacent islands toward the south and southwest were first described in geological literature by Hitchcock in 1833 (p. 143-144) as a part of the "diluvium" resulting from "the flood," although he evidently had grave doubts about the adequacy of that explanation of their origin. In 1879, Upham (1879, p. 81-92, 197-209) described many of the features discussed in this paper as a part of the more northerly of two glacial moraines that he had traced eastward and northeastward from Long Island. Commenting upon Upham's work, T. C. Chamberlin (1883, p. 377-379) stressed the lobate character of the ice sheet from which these moraines had been deposited and inferred that the moraines of Cape Cod marked a recessional stage of the ice that had retreated not long before from its terminal position on Martha's Vineyard and Nantucket.

Shaler (1881, p. 57-58; 1889, p. 46; 1898, p. 497-593) likewise recognized the glacial character of the Cape Cod landscape and described many of the features in the western part of Cape Cod as portions of a frontal moraine and its marginal outwash apron. In 1906, Wilson (p. 4, 7, 48-50) presented an argument for the "probable existence of a

great Newfoundland ice-sheet which was the source of the ice on Nantucket and on Cape Cod" and reported an "interlobate moraine extending in a southerly direction from the neighborhood of Plymouth and

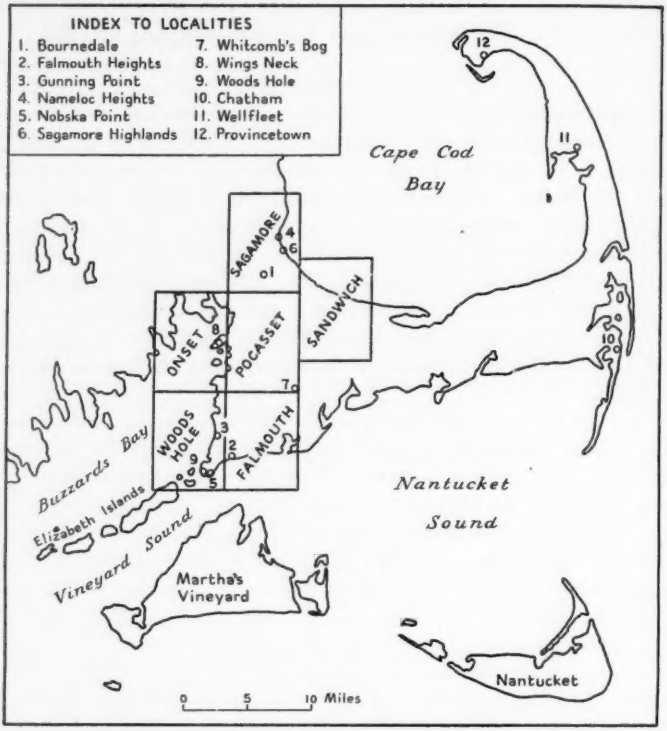


FIGURE 1.—*Outline map of southeastern Massachusetts*
Including Cape Cod. Quadrangles studied in detail are indicated by name.

Manomet Hill to Wood's Hole." He described the portion of the "Cape Cod moraine" parallel to Buzzards Bay as partly interlobate and partly marginal to the "Long Island lobe" of ice at its second stage following retreat from the northwest side of Martha's Vineyard.

During the same decade, Woodworth, Fuller, Clapp, and others were investigating the glacial features of Long Island, Rhode Island, and Massachusetts, and frequent references were made to the possible relationships between the moraines of Cape Cod and those of other localities.

In 1925, Johnson (p. 105-117) presented a convenient synthesis of the ideas thus gained and identified the moraines with which the present paper deals as a part of the "Harbor Hill moraine" extending eastward from Orient Point on Long Island. Although Woodworth's field studies were completed in 1919, his results were not published until 1934, in a memoir that embodies the most detailed and inclusive report on the geology of Cape Cod thus far available. In this report all the morainic deposits south and southeast of the Cape Cod canal in the area now under consideration are grouped together as the "Falmouth moraine," formed during late Wisconsin time after the ice margin had retreated northward from the outer moraine on Martha's Vineyard and Nantucket.

More recently, Bryan (1932, p. 176) directed attention to the "weathering, wind-worn stones and intense frost heaving . . . characteristic of many localities of southeastern Massachusetts" and suggested that "Martha's Vineyard, parts of Cape Cod and of the mainland of Massachusetts were, on this evidence, free of ice in the late Wisconsin and perhaps even through the whole of Wisconsin time." Moreover, Sayles (1939, p. 1931-1932), working principally on the more easterly portion of the Cape, observed phenomena indicative of an even greater complexity of glacial and associated deposits than had been inferred previously. It therefore seemed desirable to take advantage of the recent completion of detailed topographic surveys of the region adjacent to Buzzards Bay and to re-examine this critical area in the light of modern concepts of Pleistocene glacial geology.

PART I. GLACIAL MORAINES IN THE VICINITY OF BUZZARDS BAY

BUZZARDS BAY MORAINE

As indicated in Figure 1, the quadrangles surveyed during the 1939 field season cover the broad base of Cape Cod, nearly separated from the mainland of Massachusetts by Buzzards Bay, and the "spur" extending southwestward toward the Elizabeth Islands. The most conspicuous topographic features are the hills and dales, knobs and hollows of the boulder-strewn belts designated in this paper (Fig. 2) as the Buzzards Bay moraine and the Sandwich moraine. These moraines are connected but diverging portions of the extensive compound, morainal belt stretching from the westernmost of the Elizabeth Islands to the eastern shore of the "fore-arm" of Cape Cod, described by Woodworth as the "Falmouth moraine." The Buzzards Bay moraine was formed by the Buzzards Bay lobe of the Labrador ice sheet; the Sandwich moraine by the Cape Cod Bay lobe of the same great glacier.

The Buzzards Bay moraine extends in almost a straight line from the north-central part of the Pocasset quadrangle southward and

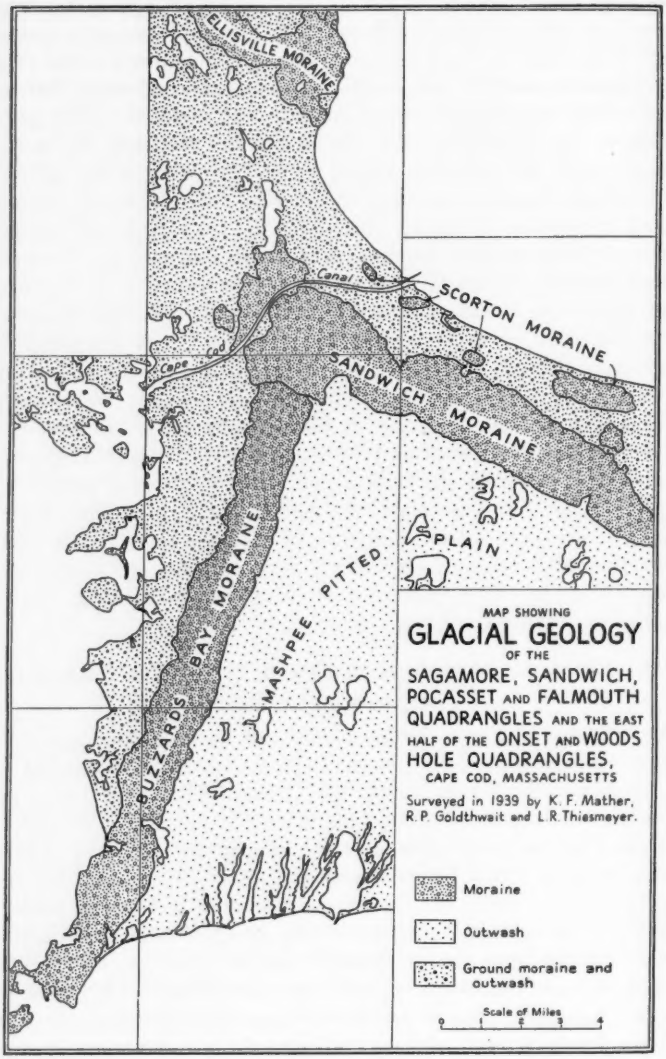


FIGURE 2.—*Map showing glacial geology of western Cape Cod*

southwestward across the northwest corner of the Falmouth quadrangle to Woods Hole, near the southeast corner of the Woods Hole quadrangle. Thence it presumably curves westward in the Elizabeth Islands, which the writers have not as yet studied. Its eastern margin is marked nearly everywhere by an abrupt change in topography and lithology, from the rough, hilly landscape of the moraine, liberally strewn with large erratic blocks and boulders, to the flat, gently sloping surface of the Mashpee pitted plain with its many deep kettle holes and its cobblestones, gravel, and sand. The western margin is much less definite; here there are gentler and more irregular slopes from the higher ground of the moraine to a gravel-strewn area with an occasional patch of till, leading down to the shore of the bay.

An observer driving southward on the highway connecting Buzzards Bay and Falmouth (State Highway 28, Fig. 3), from a point about one mile south of Bourne Bridge, sees the knobby, ridgelike moraine forming the skyline toward the east. At first the margin of the moraine is half a mile from the highway, but the two converge and, for several miles southward from the crossing of the Pocasset-Forestdale road, the highway follows rather closely the morainic boundary. South of West Falmouth, however, the highway crosses obliquely through the moraine to Falmouth, just beyond the eastern front of the moraine on the Mashpee pitted plain. Similarly the Buzzards Bay shore line and the moraine converge southward, so that in the Saconesset Hills of east-central Woods Hole quadrangle the moraine rises directly from the narrow sandy beach and continues thence to Woods Hole along the shore.

Many of the irregular hills in the northern part of the moraine rise to altitudes of 250 to 300 feet; the highest, Signal Hill, is 306 feet. The bottoms of the generally undrained depressions scattered irregularly among these hills are at altitudes of only 130 to 160 feet, thus giving a local relief of approximately 125 feet. Here the eastern face of the moraine is a somewhat broken scarp, 30 or 40 feet high, that rises abruptly from the surface of the Mashpee pitted plain at altitudes of 200 to 220 feet. In contrast the western edge of the moraine is represented by a much more sinuous line that generally falls between the 100- and the 120-foot contour line. Six or eight miles southward, near the southwest corner of the Pocasset quadrangle, the higher morainic knobs reach only 150 to 190 feet above the sea, the foot of the east-facing scarp has descended to an altitude of scarcely more than 100 feet, and the western edge of the moraine is generally between 40 and 60 feet above sea level. Nevertheless, the local relief within the morainic belt is just as great as it is farther north; depressions between the hills descend to altitudes of only 40 to 70 feet, and the water surface of

Turtle Pond, in the deepest of the holes in this portion of the moraine, is only 22 feet above sea level. Still farther southward, in the Woods Hole quadrangle, the moraine forms a peninsula jutting southwestward between Buzzards Bay and Vineyard Sound. Here its higher hills have summit altitudes between 100 and 125 feet, and the local relief is only slightly less than elsewhere. Here also the gently sloping surface of the Mashpee pitted plain beneath the eastern front of the moraine descends from an altitude of about 40 feet at Falmouth and disappears beneath the sea a mile northeast of Nobska Point.

Throughout the entire length of the Buzzards Bay moraine, from the vicinity of the canal to Woods Hole, the surface is abundantly covered with huge boulders and subangular blocks of rock ranging up to 18 or 20 feet in greatest diameter. Hardly a 5-acre patch lacks an erratic that is not at least 8 feet across. No stones of similar size are present on the pitted plain to the east where the largest rocks at most places are cobblestones or in some places are boulders 2½ to 3 feet in diameter. Moreover, on the lower ground to the west, between the moraine and the bay shore, large erratics are relatively scarce and are apparently confined to certain, sparsely scattered areas, each 50 acres or so in extent, separated from each other by hundreds of acres devoid of rocks larger than a cobblestone.

Except for these conspicuous blocks and boulders the materials composing the Buzzards Bay moraine are almost completely concealed by a dense growth of grass, shrubbery, and trees. Exposures in road cuts and gravel pits must be amplified by the liberal use of mattock, hoe, and soil auger, if one is to secure a true concept of the composition of this moraine. The casual observer would note the many pits, dug in stratified sand and gravel, and would be tempted to conclude that such materials make up the bulk of the hilly zone. Nearly everywhere, however, there is a veneer of till, 3 feet or more thick, that must be removed before the stratified sand and gravel are exposed, and at many places exposures of till are 10 to 14 feet thick. Nearly all the pits excavated in the morainic belt were intended to secure gravel and sand; consequently, where thick till was encountered, digging was promptly abandoned, and the hillside scars have long since been healed by slope wash, frost action, and the growth of vegetation.

One of the best exposures of till in the Buzzards Bay moraine was noted in a recently worked pit in West Falmouth (northeast corner of Woods Hole quadrangle), on the north side of Blacksmith Shop road half a mile east of the state highway. During the summer of 1939 this showed a fresh cut, 8 to 10 feet high, entirely in till. The middle and lower portions consist of firmly compacted gray sandy

till containing many subangular blocks and boulders that range up to 3 feet in longest dimension. There was a suggestion of irregular lamination or pseudo bedding as though ice had pressed heavily upon the deposits. The upper portion is rusty brown as a result of oxidation by ground water and is much less resistant to erosion and to excavating tools than that beneath. At one place the oxidized zone is 4 feet thick, where the large roots of a tree have penetrated far into the ground, but elsewhere the depth of oxidation is approximately 2 feet, beneath the few inches of soil.

Such compact till is one of the materials referred to as "hardpan" by the local residents and well-diggers. It is not, however, typical of the Buzzards Bay moraine. At the majority of exposures, even beneath the zone of oxidized material, the till is loose and easily crumbled. This is due to the small amount of clay in its characteristically sandy matrix or to the lack of compression by overriding ice. In general, the glacialist familiar with the drift of the Mississippi Valley and New England would refer to the nonstratified material ordinarily observed in this moraine as "sandy till" rather than as "boulder clay."

The combination of topographic features, distribution of large erratics, and scattered exposures of fairly thick bodies of till leads to the conclusion that this is a true glacial moraine, not a "pseudo-moraine" as that term is used by some glacialists. It marks a former position of the margin of the ice sheet and therefore may be appropriately referred to as a frontal moraine.

Some of the depressions in this moraine are doubtless due to the tardy melting of buried ice blocks. Others presumably result from the irregularity of the surface on which the ice was resting. The great majority, however, are irregular hollows where the total thickness of glacial debris was less than that composing the surrounding hills. A few of these hollows have been partly filled by glacio-fluvial sand and silt, and presumably there has been at least a little slope wash and soil creep in all of them. Many contain bogs, several sufficiently large and suitably located to be used for the cultivation of cranberries. In some there is a considerable accumulation of peat. A very few of the depressions near the western edge of the moraine contain permanent ponds; subsurface drainage is apparently competent to remove the rain water from any hollow, the bottom of which is not less than 50 feet above sea level.

One of the larger of the shallow, flat-floored depressions, apparently the site of considerable glacio-fluvial or glacio-lacustrine deposition, is located three-eighths of a mile southeast of Signal Hill, near the northern end of the Buzzards Bay moraine, 1½ miles northwest of the

center of the Pocasset quadrangle. Here a conspicuous treeless glade at an altitude of 200 feet is traversed by Signal Hill road and surrounded by irregular wooded hills that rise 50 feet or more above the tiny plain. The soil auger revealed that beneath a surface zone of chocolate-brown to black sandy soil, 10 inches thick, there is a zone of rusty sand 2 feet thick, and beneath that 15 feet of clean, yellow, medium to coarse sand with a few layers containing pebbles that range up to 2 inches in diameter.

In contrast with such places where the sand and gravel presumably overlie till, there are many localities where the relations of till overlying silt, sand, and gravel are clearly displayed. The walls of a recently excavated pit on the south side of the Pocasset-Forestdale road, one mile east of the state highway, near the center of the Pocasset quadrangle, were free of slumped material during the summer of 1939. Beneath the soil zone, 4 or 5 inches thick, with whitish, bleached humus at its base, there is 18 to 33 inches of light-tan, oxidized till containing many stones 4 inches or more in diameter and subangular blocks a foot or more in length. This till rests upon laminated silt or fine sand with many layers of sticky clay, containing a few, presumably ice-rafted, cobblestones and pebbles. The uppermost 4 to 10 inches of this water-laid deposit is oxidized to a bright or rusty red; the rest is gray or brownish gray. The upper 2 to 5 feet of this stratified material is contorted into folds, some of which are overturned and broken by thrust faults, as though the whole had been shoved from the west by overriding ice. An auger hole sunk in the floor of the pit, beginning 9 or 10 feet below the top of the waterlaid silt, sand, and clay, revealed the fact that the excavation had stopped at the bottom of this type of material. A few inches below the pit floor the auger encountered gray till, not quite so compact as that described in a preceding paragraph as seen in the lower part of the pit near the Blacksmith Shop road, but otherwise similar to it, and continued in till to a depth of a little more than 3 feet. Beneath the till was another zone, less than a foot thick, of laminated silt and clay, resting upon sandy gravel or sandy till into which the auger could be driven to a depth of only 2 feet.

At this locality there seems to be no valid reason for postulating any long interval of exposure of the laminated silt and clay to the agents of weathering and oxidation prior to the deposition of the overlying till. The observed phenomena may well have resulted from oxidation proceeding from the surface downward through the till to affect the uppermost few inches of the stratified materials after the final withdrawal of the ice. If that interpretation is correct, the whole deposit of intercalated till and glacio-fluvial or glacio-lacustrine mate-

rials may well have been laid down close to the edge of a slightly fluctuating body of ice, the margin of which at times was a fraction of a mile to the west and at other times advanced a similar distance to the east. In the last such oscillatory advance, the moving ice distorted the laminated silt, sand, and clay over which it rode. It is precisely the sort of heterogeneous deposit that one would expect to see in a frontal moraine constructed beneath the edge of a glacier, the margin of which shifted from time to time across a belt a mile or so wide.

TABLE 1.—*Mechanical analysis of material from Brick Kiln Road pit*

(Mm.)	(Per cent)
0.3 and over	0.2
0.25–0.29	0.3
0.15–0.24	0.1
0.10–0.14	0.7
0.04–0.09	20.0
0.02–0.03	20.0
0.01–0.02	24.0
Less than 0.01	34.0

The till and associated water-laid materials are also exposed in the "claypit" on the north side of Brick Kiln road, three quarters of a mile east of the state highway, just within the west margin of the Falmouth quadrangle. Here there is an irregular area of flat land, about 120 feet above sea level, covering 40 or 50 acres, approximately midway between the east and west boundaries of the Buzzards Bay moraine. At several points the edges of this small plain overlook deep hollows, and at other places it abuts against the slopes of irregularly rounded hills that rise 20 or 30 feet above it. The broad shallow pits in the western part of the flat area were dug to secure "clay" for the construction of tennis courts for which the material seems to be peculiarly suitable. Apparently it was also used, years ago, for the manufacture of brick.

The "clay" is in reality a finely laminated silt with alternating layers of reddish-brown and brownish-gray sediment that in some places recur so regularly as to resemble varves. Some of the grayish laminae have the fine texture and sticky consistency of clay, whereas the reddish laminae are coarser and less plastic. Individual layers are commonly only a few millimeters thick. A mechanical analysis (Table 1) of a sample taken 6 feet below the top of the deposit and containing material of both types of laminae was made by Arthur S. Knox.

Thus it would appear that approximately two-thirds of the particles range between 0.01 and 0.10 of a millimeter in size, whereas about one-third of them are less than 0.01 of a millimeter in diameter.

At both the north and south ends of the excavation the laminated silt is overlain by till, 2 or 3 feet thick and containing the customary assortment of cobblestones and blocks ranging up to 2 feet in length. Several larger erratics are scattered over the floor of the abandoned portion of the pit, and others appear on the slopes rising beyond the flat area. Where overlain by till the upper foot or two of the laminated silt is slightly arched into sinuous folds of varying dimensions; at only one place, as exposed in the summer of 1939, were these folds sheared or mangled (Pl. 1, figs. 1, 2). In general the uppermost 2 to 2½ feet of the silt is coarser, sandier, less finely laminated, and more irregularly bedded than the lower portion. Surface oxidation has proceeded irregularly downward from the base of the soil zone to depths of 2 or 3 feet and is much more conspicuous in the coarser laminae than in the fine. At certain points, near the north end of the pit, the lower 2 to 5 inches of the overlying till, where the surface of the silt is lower than on either side, is thoroughly oxidized to a deep red.

At one point, also near the north end of the pit, as developed during the summer of 1939, a glacial erratic more than a foot long was embedded 2½ feet below the surface of the laminated silt (Pl. 1, fig. 3). The laminae were bent downward beneath the erratic, conforming to its lower surface, but were horizontal a few inches above its upper surface. It was unquestionably a berg-rafted stone dropped to the floor of the pond in which the silts were accumulating at the time of their deposition. Probably some of the erratics left on the floor of the pit during excavation had a similar origin.

The maximum thickness of the exposed silt was 8 feet. An auger hole in the floor of the pit at that spot penetrated 7 feet more of the same sort of material and continued downward through 5 feet of clean, well-washed sand and gravel. The vertical section of the material at that point may therefore be given as follows:

DEPOSITS AT BRICK KILN ROAD, NORTH OF FALMOUTH

Soil	4 inches
Till; sandy, brownish	2 to 3 feet
Silt; laminated, contains berg-rafted boulders	15 feet
Sand and gravel	5 feet +

Another auger hole near the center of the flat area east of the "claypit" revealed the following sequence of materials:

DEPOSITS NEAR BRICK KILN ROAD PIT, NORTH OF FALMOUTH

Soil, with humus and roots	6 to 8 inches
Soil; rusty and sandy	1 foot
Silt and clay; laminated like that exposed in the pit, 300 yards to the west; dark, rusty sand streaks at about 8 and 10 feet. Boring stopped by pebbles in rusty sand at 11.5 feet	10 feet

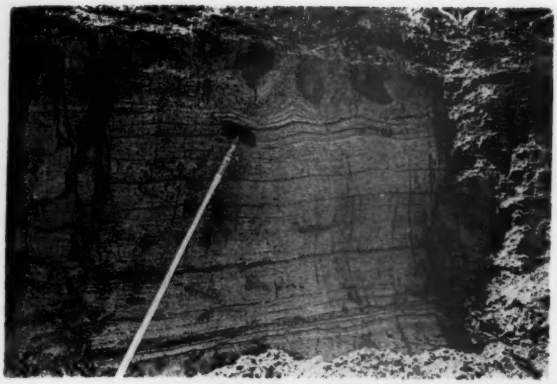


FIGURE 1. LAMINATED SILT EXPOSED IN SUMMER OF 1939 AT SOUTH END OF PIT

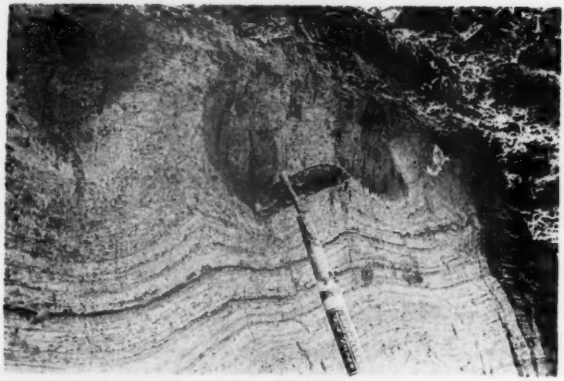


FIGURE 2. UPPER PORTION OF SAME EXPOSURE AS IN FIGURE 1 Showing lumps of till pressed into contorted laminae of silt, sand, and fine gravel.

FIGURE 3. BERG-RAFTED BOULDER APPROXIMATELY 20 INCHES LONG Embedded in laminated silt and clay at north end of pit, as developed in Summer of 1939.

GLACIO-LACUSTRINE DEPOSITS IN BRICK KILN ROAD PIT

FIGURE 1. COARSE GRAVEL EXPOSED IN LAWRENCE COMPANY PIT West of Sols Pond in Falmouth.

FIGURE 2. LARGE CLUMPS OF TILL EMBEDDED IN STRATIFIED SAND AND GRAVEL
In this exposure at southeast side of gravel pit near Teaticket, both the till clump in middle of vertical face and the one directly above the man are irregularly stained with iron oxides.

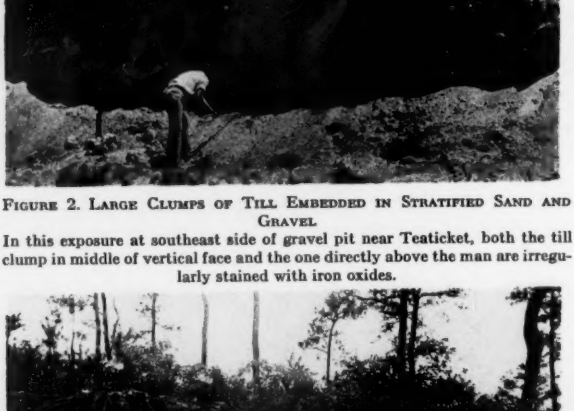


FIGURE 3. HORIZONTALLY BEDDED SAND AND GRAVEL IN ROAD CUT ½ MILE SOUTHEAST OF JENKINS POND ON ASHAMET-GRASSY POND ROAD
Bedding is truncated by gently sloping wall of a furrow trending nearly at right angles to road cut.

MATERIALS COMPOSING MASHPEE PITTED PLAIN

Samples of the laminated silt were collected in the pit at depths of 4, 6, and 8 feet below the overlying till by Arthur S. Knox for a study of the microflora which they might contain. He reports that each of the three samples rated 2,21d on the Ridgway color scale and contained unrecognizable carbonized fragments that might be remains of seaweed and unidentifiable yellow resinous bodies. No pollen was observed, but all three contained the fossils listed below, although not all the varieties of sponge spicules were noted in each:

FOSSILS IN SILT FROM BRICK KILN ROAD PIT
(identified by Arthur S. Knox)

Diatoms—very rare (marine or brackish water)
- *Melosira sulcata* (Ehrenberg) Kützing
- *Coscinodiscus radiatus* Ehrenberg

Sponge spicules—mostly broken (marine or fresh water)
- Oxea megasclera, 204 to 306 by 11.9 to 17 microns
- Styli monactinal megasclera, entirely spined, 230 by 11.9 microns
- Styli monactinal megasclera, smooth, up to 510 by 17 microns
- Fusiform acerate megasclera
- Clavate megasclera
- Large spicules, probably triaene tetraxonids (plagiotriaene), up to 102 microns in diameter
- Sterraster microsclera (fragments)

The occurrence of marine or brackish-water diatoms and marine or fresh-water sponges, all of which are Pleistocene or Recent forms, in these sediments at an altitude of more than 100 feet, and beneath the glacial till that veneers the surface of the Buzzards Bay moraine, makes it necessary to reconsider the views of Shaler concerning this major feature of Cape Cod geology. He believed (1898, p. 554-559) the ridge-like appearance of that portion of the "Cape Cod moraine" here designated as the Buzzards Bay moraine was not a result of glacial deposition but that the glacial debris was merely a thin veneer resting "on the summit of a pre-Glacial divide which separated the waters of the old Buzzards Bay river from that which formed the valley that is now Vineyard Sound." Although there is no evidence that Shaler had seen the laminated silts at or near the locality just described in detail, those sediments would undoubtedly fall within the category of "pre-Glacial rocks" described by him as composing "the Falmouth continuation of the Plymouth ridge."

The argument as Shaler might present it in the light of present knowledge would be somewhat as follows: During an early Wisconsin substage of glaciation, or better during a pre-Wisconsin Pleistocene glacial stage, an ice margin stood somewhere to the north of the Falmouth quadrangle, the whole of which was at that time much lower in relation to sea level than it is today. Sand, silt, and clay accumulated on the

sea floor, possibly in an estuarine re-entrant of the sea, and berg-rafted boulders were dropped into the laminated sediments. Diatoms and sponges, inhabiting the marine or brackish water at this place, contributed the fossils that have been discovered. Later uplift raised the sea floor at least 120 feet above sea level at this spot. Subsequently erosion removed much of the marine strata and left a high hill or ridge on the divide between the two "pre-Glacial" valleys that he mentions. Later still, the ice of some one of the Wisconsin substages advanced on the hill without greatly changing its topography and on retreat left it veneered with a thin irregular coating of till. Two alternatives present themselves, the first of which probably, though not certainly, would be approved by Shaler. (1) The marine silts might be regarded as much older than the moraine along the northwest shore of Martha's Vineyard, and therefore the hill was overridden by ice that reached that far beyond it toward the south; (2) the marine silts might be interpreted as formed after the ice margin had withdrawn far to the north from Martha's Vineyard and that therefore the interval of time between the deposition of the moraine on Martha's Vineyard and the glacial readvance to the Buzzards Bay moraine was long enough to permit a major shift in the strand line and a very considerable amount of subaerial erosion.

Neither of these hypotheses seems tenable. If the first is accepted, the deep hollows in the Buzzards Bay moraine in close proximity to the laminated silts would imply that a flat-topped hill rising steeply 50 feet or more above its surroundings and composed of unconsolidated silt and sand had withstood erosion by ice which overrode it and advanced 15 or 20 miles beyond it. If the second hypothesis is adopted, the close similarity in amount of weathering, oxidation by ground water, and modification by stream erosion that characterize the two moraines in question would seem to indicate that the time between their deposition was altogether inadequate for the large amount of subaerial erosion required.

A third hypothesis might be constructed on the theory that the ice advancing to the site of the Buzzards Bay moraine shoved the laminated silts forward and upward from their original position to their present location. This would make unnecessary any appeal to an unreasonably long interval of erosion between the deposition of the two moraines. It is, however, opposed by the fact that the silty beds appear to be quite horizontal throughout a considerable area, much too extensive to be interpreted as the flat summit of a broad anticlinal fold or as part of an overthrust block composed of such incoherent material.

Such considerations lead to the belief that the laminated silt and underlying sand were originally deposited during the construction of the moraine of which they appear to be a part. Melt water from the

ice responsible for the Buzzards Bay moraine was locally ponded in a hollow close to the irregularly frayed margin of the ice, perhaps with tongues or blocks of stagnant ice on either side. Here the sand and silt transported by the melt water filled the depression and built a small but ordinary "sand plain." Icebergs drifting across the pond dropped berg-rafted boulders in the midst of the accumulating sediment. Fossil remains of diatoms and sponges that had lived in some previous interglacial interval in the waters then occupying the present site of Massachusetts Bay to the north were transported in the ice, with the marine clay in which they were embedded, from some place where the advancing glacier plowed into the sediments on the shallow sea floor. Liberated when the ice melted, these fossils were washed into the pond by the melt water along with the particles of silt and clay, just like any other glacial erratics. Later a slight readvance of the ice shoved its thin margin over the west edge of the sand plain, disturbing the uppermost layers of silt and at last depositing the thin veneer of till that overlies the silt in the vicinity of the "claypit." Final disappearance of the stagnant blocks and tongues of ice, beneath or beyond the active ice that participated in that final movement, left the deep holes near the silt plain and the ice-contact-slope along its northeastern margin.

That the ice responsible for the Buzzards Bay moraine moved southward along a route involving the traverse of a portion of the area now occupied by the waters of Massachusetts Bay is indicated by the variety of stones in the till of that moraine. During what stages of the Pleistocene the portion of the continent immediately east of Boston was submerged and for how long the submergences continued are unknown; but the presence of occasional diatoms and fragmentary sponge spicules in the till at several places in Cape Cod, east of the Buzzards Bay moraine, indicates that the ice had access to marine clay containing such fossils and that some of them survived the vicissitudes of transport in the ice.

In the absence of bedrock suitable for the recording of glacial movement, the only basis for discovering the direction in which the ice advanced is that provided by the identification of the transported stones in terms of their known sources in eastern and northern New England. "Stone counts" were therefore made at a dozen localities in various parts of the Buzzards Bay moraine. At each exposure generally about 200 (at some places only 100) pebbles, between 1 and 3 inches in diameter, were picked up at random, an effort being made to collect all such stones that were available within a radius of a foot or two from some selected point in the midst of a body of till. All the stones in each collection were later classified on the basis of their lithologic

TABLE 2.—*Stone-count data*

Composition (in percent) of pebbles in Cape Cod moraines

Locality*	Buzzards Bay moraine								Sandwich moraine							Area of overlap				
	G-80	G-79	G-33	G-38	G-78	G-14	G-82	Average	G-55	G-42	T-80	G 98	T-82	T-83	Average	T-67	T-124	T-121	T-122	Average
Granite (including gneissoid granite, aplite, pegmatite)	69	72	70	67	61	66	58	66	55	36	38	35	57	50	45	56	56	71	58	60
Quartz-feldspar gneiss	0	6	0	1	0	0	0	1	0	5	2	4	5	3	3	0	0	0	0	0
Diorite and gabbro (including biotite gneiss)	3	1	0	0	9	9	2	3	4	4	7	3	1	2	3.5	1	4	0	3	2
Basalt (including basalt porphyry)	3	2	3	2	3	4	3	3	4	11	11	13	5	13	9.5	6	6	5	3	5
Other volcanics (mostly rhyolite and pyroclastics)	2	2	0	6	6	5	8	4	10	18	16	19	9	10	13.5	15	11	5	12	11
Schist (including greenstone)	7	7	0	3	1	1	2	3	0	3	4	5	9	9	5	3	3	7	5	4.5
Slate (including argillite)	0	0	0	1	0	0	0	t	2	3	7	4	3	6	4	0	4	2	2	2
Quartzite	12	6	18	12	15	5	11	11	14	5	10	10	4	2	7.5	9	8	6	14	9
Vein quartz	5	4	9	8	6	7	16	8	11	11	4	6	6	6	7	7	9	6	4	6.5
Arkose and sandstone	0	0	0	0	0	0	0	0	0	3	1	2	0	0	1	1	0	0	0	t
Conglomerate	1	0	0	0	0	0	1	t	0	0	0	0	0	0	0	0	0	0	0	0
Miscellaneous	0	1	0	1	0	3	1	1	1	3	2	0	1	0	1	2	0	0	1	1
	Southwest				to			Northeast	West			to			East					

* Localities designated G represent collections by Goldthwait; T, Thiesmeyer.

character. The results obtained by the more significant of these "stone counts" are listed in Table 2. At the same time, specific identification of stones derived from many of the known formations of eastern Massachusetts, southern New Hampshire, and southern Maine was attempted. Valuable assistance in making such identifications was rendered by L. W. Currier, Marland P. Billings, Harry Berman, and other geologists familiar with the geology of those areas.

By far the most abundant rocks among the erratics of this moraine are the granites and their close allies. Most of these were undoubtedly derived from the Dedham granodiorite and associated rocks of the region adjacent to the Boston basin. Others closely resemble certain of the granites exposed in southern Maine. Diorite, of the type known to occur near Salem, Massachusetts, and diabase, identified as having been derived from the widely scattered system of dikes named for Medford, are fairly common. Volcanic rocks of the types occurring at Nantasket and Lynn comprise a small but significant fraction of the assemblage. As noted by Howe (1936, p. 394-396) the fragments of the distinctive felsite known to occur in place only in the northern part of the town of Hingham, Massachusetts, are sparingly scattered throughout the Buzzards Bay moraine. On the other hand, no specimen of granite that could be identified as having been derived from Quincy granite was found. Among the few fragments of sedimentary rock are Roxbury conglomerate and Cambridge slate, the most widespread of the formations in the Boston basin. A few small fragments of laminated, nonfissile slate, like the varved slate of Squantum, Massachusetts, were also noted. Thus, it would appear that ice moving southward from southern Maine and southeastern New Hampshire across the eastern margin of Massachusetts could have gathered all the diverse materials found in the Buzzards Bay moraine.

SANDWICH MORAINE

A belt of hummocky and relatively high terrain, known to the local residents as "the backbone of the cape," stretches eastward from the Cape Cod canal across the northern part of the Pocasset and Sandwich quadrangles. Woodworth considered this a part of the Falmouth moraine, contemporaneous in origin with the other portions here designated as the Buzzards Bay moraine. It is, however, a distinct moraine, differing significantly from the southwestward-trending moraine in composition, in relations to the lobate margin of the ice sheet, and in time of emplacement. The name Sandwich moraine is therefore applied to it, inasmuch as it traverses for many miles the northern part of the town bearing that name.

* Localities designated G represent collections by Goldthwait; T, Thiesmeyer.

Driving east from Sagamore to Dennis on the principal highway, one has constantly on his right the irregular skyline of this "backbone" of Cape Cod. In the Sandwich quadrangle, the higher summits stand between 200 and 300 feet above sea level; of these, Telegraph Hill, with its summit altitude of 292 feet, one mile south of Sandwich village, is the highest point. Eastward along the moraine the hilltops decrease somewhat in altitude so that in the belt just south of West Barnstable the higher summits are between 150 and 200 feet above sea level. Across most of the Sandwich quadrangle the moraine is more than a mile in width but, at the eastern margin of that quadrangle, it narrows to half a mile.

The surface of this morainic belt is everywhere characterized by irregular hills and undrained depressions that give a topographic expression quite like that of the Buzzards Bay moraine. In the area extending a mile or so east and southeast from Telegraph Hill, many of the small hills are elongated parallel to the trend of the moraine. Presumably these rude ridges are the result of deposition of frontal debris or of the pushing forward of marginal drift during minor fluctuations of the ice front.

The south face of the Sandwich moraine overlooks the Mashpee pitted plain along a well-defined boundary. South of Telegraph Hill and along Popple Bottom road the moraine front appears like a high ridge strewn with scattered erratics. The moraine boundary is, however, complicated just north of the Cape Cod Airport in the southeast part of the Sandwich quadrangle by subsidiary short ridges of bedded gravel which extend obliquely into the main plain (Fig. 3). As explained in the discussion of the Mashpee pitted plain, these were probably formed near the ice edge prior to the deposition of the main moraine.

The northern part of the Sandwich moraine is very irregular, commonly sandy, and low, like the west side of the Buzzards Bay moraine. There is, however, a marked topographic break along the north boundary of the Sandwich moraine, even more conspicuous and continuous than that along the west side of the Buzzards Bay moraine. North of the Sandwich moraine lie areas covered with salt marsh, low sand plains, and a few distinct and entirely separate hills of the Scorton moraine (Fig. 2), described on a subsequent page.

At most of the exposures the material in the Sandwich moraine is sandy till. A representative pit, on State Highway Number 130, opposite the entrance to the Telegraph Hill fire tower, shows a 10-foot thickness of till, including a few big angular blocks and a few clumps of well-sorted sand. These sand clumps may be the remains of frozen masses picked up by the invading ice. Hard-packed, gray till or "hardpan," believed to be

of the same age as the loose, sandy till elsewhere in the moraine, is exposed behind the farm at the north end of Gully Lane east of Sandwich Village and at a few other places.

Although gravel pits and deep road cuts are not so numerous in the Sandwich moraine as in the Buzzards Bay moraine, there are enough such exposures to indicate that the former, like the latter, is essentially a veneer of till, 5 to 30 feet or more thick, resting on disturbed glacio-fluvial sand and gravel. Road cuts, for example near the north edge of the moraine along Quaker Meetinghouse road and Chase road, reveal sandy till underlain by bedded sand.

Large erratics are generally less numerous on this moraine than on the Buzzards Bay moraine. In the region north of Telegraph Hill or along Maple Swamp road near the center of the Sandwich quadrangle a person may travel almost a mile without seeing one. On the other hand, many areas in the southern part of the moraine are liberally sprinkled with large blocks. Along Forest Street, south of West Barnstable, and north of Popple Bottom road, northeast of Farmersville, many boulders and blocks are 2 to 20 feet long. The largest single erratic in this moraine, exclusive of the area of overlap on the Buzzards Bay moraine, may be seen one mile south of East Sandwich; it measures approximately 30 by 15 by 10 feet.

Explanation of the relative scarcity of boulders may be found in the types of rocks represented. Whereas more than 95 per cent of the large erratics in the Buzzards Bay moraine are granite or granite gneiss, from 5 to 20 per cent of the larger stones in the Sandwich moraine are composed of basalt or other volcanic rock. Granite ledges are notorious for yielding large blocks wherever they are crossed by moving ice. We may infer that as it approached Cape Cod the lobe of ice that formed the Buzzards Bay moraine passed over a larger area underlain by granite than did the ice that formed the Sandwich moraine.

The composition of the pebbles, 1 to 3 inches in diameter, also indicates a significant difference in the till of the two moraines. Fragments of bright-red, arkosic sandstone and conglomerate, finely laminated black and white quartzite, and black slate are common in the Sandwich moraine, whereas none of these rock types is found in the Buzzards Bay moraine. Moreover, the identification of the rock types represented among 200 pebbles of such sizes collected at random at each of six localities in the Sandwich moraine shows different percentages of the types that are common to both moraines. (*See* Table 2.) On the average only 45 per cent of the small stones in the Sandwich moraine are granite, and the maximum for these collections is 57 per cent, whereas in the Buzzards Bay moraine the average is 66 per cent, and the minimum is 58 per cent.

Basalt pebbles make up nearly 10 per cent of the smaller stones in the Sandwich moraine but only 3 per cent of those in the Buzzards Bay moraine. Other volcanic pebbles average 13 per cent in the Sandwich moraine and only 4 per cent in the Buzzards Bay moraine. Of course, the pebble content of any moraine changes from mile to mile of its length, but here the difference in rock content can be traced to within a few hundred feet of the boundary between the two moraines.

Doubtless, the sources from which the ice secured the materials that were deposited during the construction of the Sandwich moraine are now for the most part concealed beneath the waters of Cape Cod Bay and the Gulf of Maine. The fragments of red sandstone and conglomerate, many of which were noted in the gravel pit just west of highway 130, are particularly interesting in this connection. Lithologically, these specimens resemble the Triassic redbeds in every respect. Their most likely source is in the Triassic basin extending southwestward from the head of the Bay of Fundy and largely submerged beneath the Gulf of Maine. As interpreted by Johnson (1925, p. 295) the Triassic rocks extend along the sea floor to a point only 100 miles northwest of the Sandwich moraine.

RELATIONS BETWEEN SANDWICH AND BUZZARDS BAY MORAINES AND MASHPEE PITTED PLAIN

The Mashpee pitted plain and its origin are discussed in detail on subsequent pages, but it is desirable to summarize briefly here the conclusions that have been reached concerning the relations between the two large moraines of western Cape Cod and the adjoining outwash plain. The data now available indicate that the Mashpee pitted plain was essentially constructed by meltwater issuing from the re-entrant angle between the Buzzards Bay lobe and the Cape Cod Bay lobe of the Laurentian ice sheet at a time when its margin stood some distance northwest of the Buzzards Bay moraine front and north of the Sandwich moraine front. After the greater part of the sand and gravel of the plain had been deposited, both lobes of ice pushed forward, plowing into and overriding the previously deposited outwash. Thus the frontal moraines were deposited on the edges of the fan-shaped plain. Meltwater issuing from the ice while these two moraines were under construction contributed relatively little debris to the outwash plain.

At several places, large blocks of stagnant ice buried in the sand and gravel of the plain were overridden by the readvancing ice and remained intact until after the moraines had been constructed and the lobate margin of the ice sheet had again retreated northwestward and northward. When these buried ice masses finally melted, large kettle holes were formed, partly in the moraine and partly in the adjacent plain

(Fig. 3). Thus the large depression traversed by Forestdale road, 0.7 mile southwest of Telegraph Hill in Sandwich, extends from the Sandwich moraine into the pitted plain. A similar origin is indicated for the long, irregular kettle hole, 1.4 miles due east of Halfway Pond in the northern part of the Pocasset quadrangle, the western half of which is within the Buzzards Bay moraine, the eastern half within the Mashpee pitted plain. Likewise, Long Pond, 1.5 miles north of Falmouth, and Oyster Pond, 1.8 miles northeast of Woods Hole, occupy kettle holes partly within the Buzzards Bay moraine and partly in the Mashpee pitted plain.

OVERLAP OF SANDWICH MORAINE UPON BUZZARDS BAY MORAINE

The Sandwich and Buzzards Bay moraines coalesce in the area south of the Cape Cod canal and extending from the Sagamore into the Pocasset quadrangle. (*See* Figure 2.) The data observed here indicate that the Buzzards Bay lobe of ice had begun to retreat before the Cape Cod Bay lobe had completed its readvance to the position marked by the Sandwich moraine. Consequently the west end of that moraine overlaps the north end of the Buzzards Bay moraine.

This area of overlap is as irregular and knobby as any parts of either moraine that lie to the south or east. The morainic topography extends northwestward across the canal as far as Great Herring Pond and includes the Bournedale Hills, traversed by U. S. Highway No. 6 as it approaches the Sagamore Bridge from the west. The highest point here is Signal Hill, 211 feet in altitude. Many large granite boulders are strewn over this portion of the Sandwich moraine and lie in clusters along the transmission line on either side of the canal and in the Shawme State Forest. The largest of these, Sacrifice Rock, half a mile northwest of the fire tower on Signal Hill, has split into several parts that lie against one another, but portions of its surface retain the marks of sandblasting. Its dimensions originally were at least 30 by 20 by 15 feet.

Cuts along the transmission lines and adjacent country lanes have exposed sticky, tan to brown till with abundant buried erratics of granite and other rocks. Several pits at different places disclose stratified gravel associated with till. On Canal View road, both north and south of the boundary between the Pocasset and the Sagamore quadrangles, there are pits in coarse, stratified gravel that include water-worn boulders, a few of which are as much as 4 feet long. When the Cape Cod Canal was excavated through a low portion of this area of overlap and along the natural drainage line leading southwestward from Great Herring Pond to Buzzards Bay, it exposed sandy till with huge embedded erratics overlying stratified sand and gravel. Here as elsewhere

in the Sandwich and Buzzards Bay moraines, till was spread over earlier glacio-fluvial deposits, and some glacio-fluvial deposition accompanied its formation.

Certain types of pebbles found in the Sandwich moraine but not observed in the Buzzards Bay moraine are present in this area of overlap. These are chiefly chips of black slate, pink quartz-feldspar gneiss, red arkose, and laminated quartzite. The four pebble counts (*see* Table 2) made in this area indicate that the proportions of basalt, quartzite, and vein quartz are intermediate between those found to be characteristic of the rest of the Sandwich moraine and of the Buzzards Bay moraine. The percentage of granite pebbles resembles that in the Buzzards Bay moraine, whereas the percentage of volcanic pebbles is like that of the Sandwich moraine. This may be explained adequately as the result of overriding of the northern end of the Buzzards Bay moraine by southwestward-moving ice of the Cape Cod Bay lobe. As the ice front fluctuated back and forth in a fairly narrow belt, its debris was mixed with that into which it plowed or over which it advanced.

The southern limit of the till displaying these lithologic characteristics indicative of mixed origin is a rather definite line just south of Kendrick road near the north margin of the Pocasset quadrangle. This line also marks a notable topographic change, with higher hummocks and ridges lying to the south. Thus defined, the belt of what appears to be the westward extension of the Sandwich moraine cuts across the trend of the Buzzards Bay moraine and northwestward for a short time.

Banked against this westernmost edge of the Sandwich moraine, in the area immediately east of the traffic circle at the south end of the Bourne Bridge, there are unusually thick bodies of stratified sand and gravel, including many very large boulders. These constitute the apex of a narrow fan of outwash, spread by meltwater from the Cape Cod Bay lobe that escaped southward between the Buzzards Bay moraine and the front of the Buzzards Bay ice lobe when that ice had withdrawn 2 or 3 miles back from the moraine that it had just constructed. The nature of these deposits west of the Buzzards Bay moraine has been adequately discussed elsewhere (Mather, *et al.*, 1940, p. 40-43), and reference to them is made here merely to cite the additional evidence they provide for the conclusion that the deposition of the Sandwich moraine was completed somewhat later than that of the Buzzards Bay moraine.

SCORTON MORAINE

Between the Sandwich moraine and the shore of Cape Cod Bay there is a discontinuous line of elongate hills to which the name Scorton moraine may appropriately be applied. (*See* Figure 2.) Spaced at

unequal intervals from Sagamore Village through Sandwich to Barnstable, these hills include the hill at the northern edge of Sagamore Village, Sagamore Hill, Town Neck Hill, Spring Hill, Scorton Neck, and Scorton Hill. As a rule, each hill displays a broad, flat summit characterized by irregular hummocks and small kettle holes as much as 25 feet deep. Large erratics are sprinkled over each segment of the moraine. As in the Sandwich moraine, these are neither closely packed nor very abundant, but some are of remarkable size.

The structure of these hills is characteristically displayed in a pit near the highway at the south side and near the west end of Scorton Neck. At this locality, sandy till, 6 to 8 feet thick and including irregular masses of sand, rests on stratified sand containing thin seams of gravel. Again, on the hummocky top of Spring Hill, a mile northwest of East Sandwich, a small pit exposes compact till. The soil auger disclosed that this till rests on stratified sand a few feet below the floor of the pit. There are similar indications that Scorton Hill and the hill north of Sagamore Village are also underlain by sandy gravel and veneered with till.

The manner in which the ice deposited this discontinuous moraine is not yet clear. Further studies in adjacent areas must precede final inferences concerning the origin of the moraine and its associated deposits, but some conclusions and tentative working hypotheses can be developed from the information now available.

Most likely the Scorton moraine was never a continuous belt like the Sandwich moraine. If the ice retreated from the Sandwich moraine largely through downwasting and consequent thinning, a very ragged margin would be produced along which large masses of ice may have been quite stagnant. Between these and in the re-entrant angles along the ice front, glacio-fluvial gravel and sand would be deposited by the meltwater. Readvance of thicker portions of the ice sheet from the north would push forward the deeply indented, sawtooth edge of the Cape Cod Bay lobe of ice to override the masses of inactive, debris-clogged ice along its margin and the bodies of glacio-fluvial debris that had been deposited between them. Till would then be piled on top of the patches of stratified sand and gravel. Later, when all the ice had melted away, hills of gravel capped with till would remain where the thicker bodies of gravel had been deposited, and the lower ground near-by where stagnant ice had stood would be covered unevenly with thin till. Still later the meltwater from the dwindling ice at the north might spread sand and sandy gravel over these lower areas and blanket the till. The area would thus consist of isolated, till-veneered hills of gravel standing like islands in a sea of sand. This is essentially the condition of the Scorton moraine today.

ELLISVILLE MORAINE

A belt of very irregular topography, the surface of which stands generally higher than the adjacent countryside and is dotted with thousands of large erratics, extends northwestward across the northern portion of the Sagamore quadrangle from the shore of Cape Cod Bay near Ellisville. (*See* Figure 2.) The numerous exposures of sandy till, the many knobs and kettles, the large erratics, the linear plan of this hilly terrain, and its geographic relations to other features of the landscape indicate that it marks a temporary halt in the recession of the ice responsible for the moraines that it had earlier deposited south and east of the Cape Cod Canal. It may therefore be appropriately designated as the Ellisville moraine.

In contrast to the south and east fronts of the Sandwich and Buzzards Bay moraines, there are few places with distinct breaks in topography along the southwest margin of the Ellisville moraine. Instead, the boulder-strewn moraine merges along an irregular line with the much-pitted northern edge of a large outwash plain that extends southward and southwestward to the head of Buzzards Bay at Wareham. This may be designated as the Wareham pitted plain and displays many features quite like those of the Mashpee pitted plain. Its surface, however, rises to the higher levels of the Ellisville moraine, and in places, as at flat-topped Mountain Hill, a mile southwest of Ellisville, it stands much higher than the immediately adjacent moraine to the north or northeast. Evidently, the Ellisville moraine and the bordering Wareham pitted plain were formed simultaneously. The moraine was not superimposed on the margins of a pitted plain by readvance of ice, as were the Buzzards Bay and Sandwich moraines, but was constructed when the continuous northward retreat of the ice front was halted for a considerable interval, during which the edge of the Cape Cod Bay lobe fluctuated only within narrow limits within the area now covered by the morainic deposits. Further details concerning the Ellisville moraine and the Wareham pitted plain have been published elsewhere (Mather, Goldthwait, and Thiesmeyer, 1940, p. 46-52).

SUMMARY

Three distinct types of recessional moraines are present in the general vicinity of Buzzards Bay near the west end of Cape Cod. The Buzzards Bay and Sandwich moraines were constructed along the margins of the Buzzards Bay lobe and the Cape Cod Bay lobe of the Laurentian ice sheet during the Wisconsin glacial stage by ice that readvanced from a position north and northwest of these moraines to which it had earlier retreated. Thus the morainic debris was superimposed upon the margins

of the Mashpee pitted plain and represents in part the result of ice push as well as the deposit made by melting ice.

The Scorton moraine is a discontinuous series of till-veneered hills marking the ragged edge of the Cape Cod Bay lobe at a time of ice wastage that caused the irregular withdrawal of the ice front from the smooth line of the Sandwich moraine. Slight oscillations of the active ice within the Cape Cod Bay lobe resulted in the overriding of the stagnant ice along its frayed margin and left the moraine hills standing in the midst of glacio-fluvial deposits.

The Ellisville moraine is the least complex of recessional deposits. Continuous northward retreat of the margin of the Cape Cod Bay lobe of ice was temporarily halted. The recessional moraine and the bordering apron of glacio-fluvial debris, the Wareham pitted plain, were contemporaneously deposited.

PART II. MASHPEE PITTED PLAIN

SURFACE FEATURES AND COMPOSITION

The Mashpee pitted plain lies within the angle between the Sandwich and Buzzards Bay moraines and occupies about 100 square miles of land surface on Cape Cod (Fig. 3). Several square miles of sea floor beneath Nantucket and Vineyard Sounds represent its submerged southern margin. The plain is in essence a slightly dissected, subaerially constructed, alluvial fan with its apex at an elevation of 220 feet, close to the place of overlap of one moraine upon the other, and its surface sloping gently to the curving shore line, 10 to 14 miles distant toward the south and southeast. At the east, beyond the Pocasset and Falmouth quadrangles, it coalesces with a similar fan which according to Chute (1939, p. 12) may have developed simultaneously.

The surface of the plain is modified by two types of depressions: (1) irregular pits or kettle holes of all sizes, some of which are as much as a mile in diameter and 100 feet in depth, distributed at random, and (2) long furrows increasing in breadth from a few feet to 1000 feet and in depth from a few inches to more than 30 feet as they radiate southward and southeastward across the lower half of the plain. Woodworth (1934, p. 271) called these linear depressions "creases," and others have referred to them as "valleys," "arroyos," or "channels." The designation "furrow" is, however, adopted here because it is a noncommittal term without any implication concerning their mode of origin.

The surface slope of the entire plain, restored by filling in, in imagination, the pits and furrows, is very gently concave (Fig. 4). From the apex south and southeast for 6 miles, the average gradient is 15 to 20 feet per mile. Indeed, close to the apex some nonpitted portions of the plain

incline as much as 25 feet in one mile. The outer and lower half of the plain, on the other hand, has a surface slope of only 12 to 15 feet per mile.

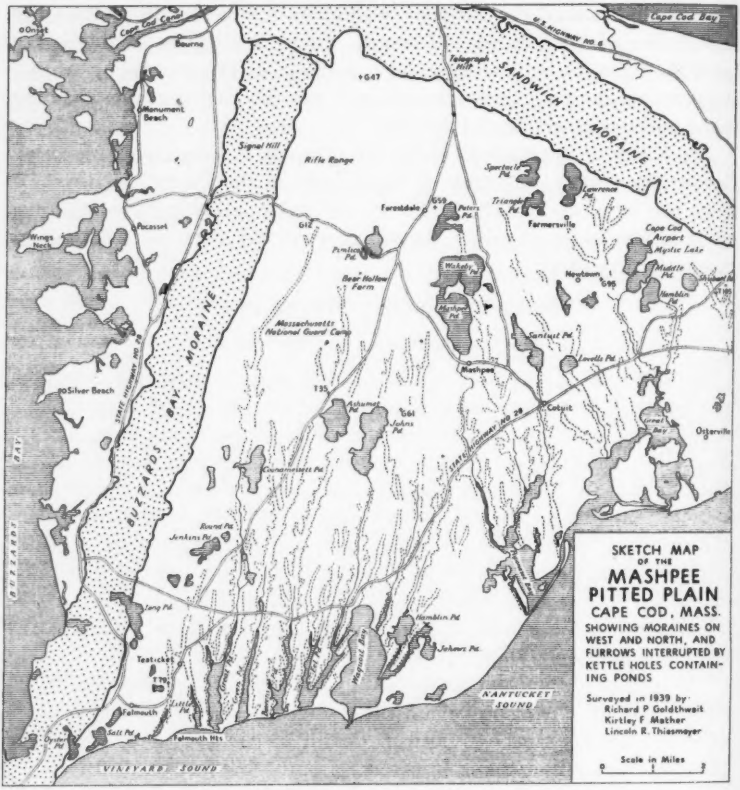


FIGURE 3.—*Sketch map of western Cape Cod*

Showing Mashpee pitted plain, bounded on north and west by Sandwich and Buzzards Bay moraines. Furrows traversing plain are delineated by broken lines. Localities at which stone counts recorded in Table 3 were made are indicated by letter and number, *e. g.*, G59.

The material underlying the plain is bedded sand and gravel, accumulated to a thickness of at least 120 feet, the depth of the deepest pits. The average components vary from well-rounded gravel near the apex to subangular sand near the distal edge. Associated with the gravel in the higher portion of the plain are abundant, small, rounded or subangular boulders, 6 to 24 inches in diameter. One pit along the Pocasset

road, 1000 feet north of a road intersection, the altitude of which is given on the topographic map of the Pocasset quadrangle as 182 feet, disclosed 4 to 6 feet of very coarse gravel and small boulders overlying 3 feet of cross-bedded sand with thin lenses of gravel. A trench in the Massachusetts National Guard Military Reservation (Camp Edwards) north of Snake Pond and many shallow road cuts in the northern part of the reservation exhibit similarly coarse gravel. In contrast, on the outer, seaward edge of the plain, the many low wave-cut cliffs almost exclusively expose only sand. Pits excavated for sand, used in surfacing roads and bogs, are scattered along the main highway leading northeastward from Falmouth and around each cranberry bog in this vicinity. Very little coarse gravel is available in these pits.

There are exceptions, however, to this gradation in coarseness. Most notable is the Lawrence Company pit, just west of Sols Pond in Falmouth. Although this locality is only 2 miles from the outer edge of the plain, it shows a 6-foot layer of very coarse gravel containing rounded boulders, some of which are as much as 18 inches in diameter (Pl. 2, fig. 1). On the floor of the pit are a few boulders as much as 30 inches long. Similar coarse gravel with rounded cobbles 8 inches in diameter occurs in a near-by pit just southwest of Teaticket. The eastward dip of the layers and slope of the surface between pits in this vicinity suggests that this abnormally coarse material was deposited by meltwater issuing from the adjacent Buzzards Bay lobe while the Buzzards Bay moraine was under construction. Apparently streams from that source built a small local fan on top of the more extensive and older deposits of the Mashpee pitted plain.

The bedding of the sand and gravel composing the fan is essentially parallel to the surface of the plain. Although cross-bedding and channelling are common, no foreset delta structure was observed. The "scour and fill" nature of the deposits is evident at many places where lenticular gravel layers lie between cross-bedded and laminated sands.

The features described above—fan shape, definite concave slope, general decrease of grain size from apex to distal edge, and beds parallel to the surface but vertically variable in texture—are characteristic of deposits of present-day anastomosing streams. The lack of foreset delta structure, even in pits at the distal edge of the exposed plain, indicates that this is a subaerial fan and was not built to accord with any lake surface or sea level.

GLACIAL ORIGIN

The very large natural pits in the plain are most convincing evidence that this is glacial outwash. They occur chiefly in groups near Falmouth, Coonamessett, the rifle ranges in the Massachusetts National Guard

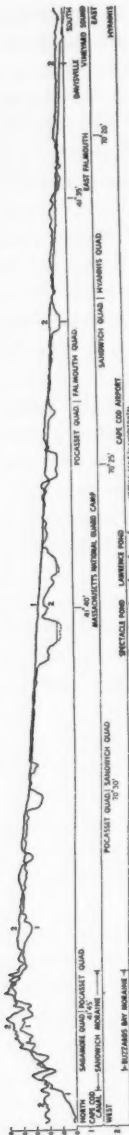


FIGURE 4.—*Profiles across Mashpee pitted plain*

Military Reservation (Camp Edwards), Farmersville, Newtown, the vicinity of Johns Pond, Waquoit, and Osterville. There is no system or regularity to the grouping or to the positions of groups of pits in the plain. More than 30 pits exceed half a mile in diameter. They range in depth from 50 to 120 feet, and several are enclosed completely by gravel walls. The most prominent examples are the depression which contains Snake Pond, west of Forestdale, and the hollows near the center of the Pocasset quadrangle. Some 500 smaller pits are shown as depression contours on the new topographic maps of the United States Geological Survey, and many others observed in the area under discussion are too small to show on maps of the scale used. The side slopes of the pits are inclined 5°, 10° and even as much as 20°. Where closely grouped, the slopes around adjacent depressions may coalesce to form saddle-shaped ridges. The pits vary greatly in shape; some are nearly circular, others elongate oval, and some quite irregular. Within an area of compound pits, the surface may resemble the hummocks and hollows of small "kame and kettle" deposits so common in valleys throughout New England.

These features—the irregular grouping, the great depth, steep slopes, and irregular shape—are prime characteristics of glacial kettle holes. Furthermore, several of these pits, such as those occupied by Long Pond and Oyster Pond near Falmouth, or the elongate depression southwest of Telegraph Hill in Sandwich, extend from the plain into the bordering moraines. As these hollows interrupt the continuous front of the moraine, they must have formed after the moraine was built on top of the outwash plain. Clearly, all these depressions must have resulted from the melting of ice that had been buried during the deposition of the plain.

Two other conditions point to the fact that this is glacial outwash: (1) the plain heads at an altitude of more than 200 feet; river deposits at this altitude could have formed here only during the existence of glacial ice to support the loaded channel of a stream at this high level; (2) fragments of rock of the types known to be in place many miles to the north and west are abundant among the pebbles in the gravel of the plain. Pebbles of the Squantum tillite member of the Roxbury conglomerate, for example, were found at Falmouth Heights and in a pit west of Coonamessett River, northeast of Round Pond. Carboniferous conglomerate pebbles from the Boston or Narragansett Basins were found at a dozen or more exposures. Gray-green melaphyr pebbles like those of the Boston region were observed at a number of localities. The scarcity of these types suggests that they are the few pieces that travelled 25 to 60 miles from their sources and became mixed with far more abundant stones from nearer sources. Varieties of rock like the

TABLE 3.—*Stone-count data*

Composition (in per cent) of pebbles in the Mashpee pitted plain on Cape Cod

Locality*	T-105	G-95	G-59	G-47	G-12	G-40	T-35	G-61	G-81	G-65	T-79	G-35	Averages: Mashpee pitted plain	Buzzards Bay moraine	Sandwich moraine	Area of overlap of moraines
Granite (including gneissoid granite, aplite, pegmatite)	61	59	71	62	67	70	71	76	75	47	68	72	67	66	45	60
Quartz-feldspar gneiss	5	2	0	0	0	0	0	0	0	1	0	2	1	1	3	0
Diorite and gabbro (including biotite gneiss)	3	3	2	6	6	2	0	0	1	5	1	2	2.5	3	3.5	2
Basalt (including basalt porphyry)	7	4	1	5	0	6	2	2	1	15	3	2	4	3	9.5	5
Other volcanics (mostly rhyolite and pyroclastics)	9	17	6	9	8	4	10	7	11	11	5	6	8.5	4	13.5	11
Schist (including greenstone)	5	2	3	2	7	2	0	1	0	3	7	2	3	3	5	4.5
Slate (including argillite)	2	2	2	1	0	1	0	0	1	0	0	0	1	t	4	2
Quartzite	5	5	7	8	2	6	7	7	5	7	9	7	6	11	7.5	9
Vein quartz	3	7	7	6	9	9	9	6	5	9	6	6	7	8	7	6.5
Arkose and sandstone	0	1	0	0	0	0	0	0	0	1	0	0	t	0	1	t
Conglomerate	0	0	0	0	0	0	0	1	1	0	0	0	t	t	0	0
Miscellaneous	1	0	1	2	1	0	1	0	1	1	1	1	1	1	1	1
	East to West						North to South									

* Localities designated T represent collections by Thiesmeyer; G. Goldthwait.

Dedham granodiorite, which make up a large part of every collection of pebbles (*See* Table 3), probably came from buried ledges of bedrock only a few miles away, although no outcrops are visible within 15 miles of the plain. Regardless of the distance they may have travelled, the only plausible sources for these types of rock are north or west of Cape Cod. Southward or eastward transportation by moving ice and its associated melt-water is implied by their presence in the deposits described here.

RELATIONS TO ICE FRONTS

As outlined in preceding paragraphs the Buzzards Bay and Sandwich moraines were deposited by readvance of the ice after construction of the Mashpee pitted plain and overlap the stream-washed sand and gravel at its western and northern margins. The gravel underlying the till of those moraines is in all respects like that of the pitted plain, but its surface is 50 to 100 feet lower. As may be seen along the western side of the Buzzards Bay moraine, however, the surface of the bedded deposits beneath the till is not smooth like much of the plain, and in places large angular blocks are embedded. Therefore, it is believed that these portions of the plain, now buried under the moraine, were originally at the margin of the great fan. There the outwash was deposited on and around irregular masses of ice at the ragged edge of the ice sheet. Shifting currents, such as would result from movements of water over this rough and changing surface, deposited layers sloping at unsystematic angles. The irregularity of the surface increased as the buried ice melted. From the ice itself angular blocks fell and slid into the gravel at some places.

The water that built the pitted plain carried fragments of rock like those found in both the Sandwich moraine on the north and the Buzzards Bay moraine on the west. Counts of the number of different kinds of rocks among 400 pebbles selected at random at each of the 12 locations bear out the more casual observations made at innumerable pits. (*See* Table 3.) Only one stone count, from west of Peters Pond, is closely similar to counts in the Sandwich moraine, and only two stone counts—one west of Massachusetts National Guard Military Reservation (Camp Edwards) on Turpentine Road, the other east of Shallow Pond on Mill Road—are much like counts in the Buzzards Bay moraine. The average of all pebble counts from the plain shows a proportion of granites, diorites, and basalts like that of the Buzzards Bay moraine, whereas the proportion of other volcanic and quartzite pebbles is like that of the Sandwich moraine. As the types of rock contained in the two moraines are significantly different, it would appear that the water that bore gravel to the pitted plain gathered material

* Localities designated T represent collections by Thiesmeyer; G, Goldthwait.

from sources within both lobes of the ice sheet. During transportation these pebbles were fairly well commingled.

The uniform slope of the plain toward the south and southeast, away from the apex but parallel to each moraine front, shows that the water entered the plain somewhere near or at the present apex, rather than here and there along the line of the moraines. At only two localities (just southeast of Signal Hill in Bourne and just northwest of Falmouth) does the plain show a marked slope away from the bordering moraine. In Bourne this deposit is made of sand and is shaped like a small fan. In Falmouth it is made of coarse gravel already described in the discussion of the Lawrence pit. Presumably, water flowing through breaks in the ridge on the east side of the Buzzards Bay moraine or directly from the face of the Buzzards Bay glacial lobe superimposed fans on top of the main plain at these two localities.

At two other places constructional gravel deposits rise above the general surface of the pitted plain. The more conspicuous one—Falmouth Heights, on the shore of Vineyard Sound—is elliptical, and its top is 35 feet above the adjacent plain. On the seaward side a 40-foot wave-cut cliff consists entirely of cross-bedded sand and gravel. In the upper layers of the gravel are many ventifacts. This gravel must have been deposited as outwash from the ice prior to the building of the pitted plain, for its altitude implies ice or other gravel immediately surrounding the present site to enable streams to deposit at such a high level. This deposition occurred before any other event registered by the surface deposits on this end of Cape Cod. Presumably it occurred while the ice front was retreating from Martha's Vineyard toward the line of the Buzzards Bay and Sandwich moraines.

Originally the deposit may have been far more extensive than it is now. If so, most of it must have been removed by the braided streams that built the outwash fan around it and by ocean waves now active at present sea level. It is more likely, however, that this body of gravel never extended very much farther toward the north than it does today and that it represents the apex of an outwash fan built southward from the ice front when the latter was temporarily standing close by the Heights. Rather rapid recession of the ice then removed the support from the north side of the fan, and subsequent outwash deposits were never built to so great an altitude.

The other outwash deposit above the pitted plain is a discontinuous, broad, low ridge trending southeast from the edge of the Sandwich moraine, north of the Cape Cod Airport, out into the plain as far as the Barnstable-Falmouth road. At its northwest end, this ridge blends into the moraine front; at its southeast end, it is half a mile south of the

moraine. Its undulating crest rises 10 to 25 feet above the adjacent plain, and it consists of well-bedded and cross-bedded sandy gravel without boulders. The gentle south slope and slightly steeper north side of the ridge suggest that it, too, was deposited by glacially fed streams when the ice edge stood close to or along its northern margin.

TILL CLUMPS

Masses of unmodified till are not ordinarily found in glacial outwash at any great distance from the moraines with which the outwash is associated. Irregular clumps of unmistakable till were observed, however, at six localities in the Mashpee pitted plain, far out on the glaciofluvial fan. Five of these occurrences were in pits excavated for sand or gravel: (1) immediately west and (2) a mile north of camp headquarters in the Massachusetts National Guard Military Reservation (Camp Edwards); (3) in the northern part of Bear Hollow Farm south of Snake Pond; (4) a mile west of Bear Hollow Farm; and (5) south of the cranberry bogs in Teaticket. The sixth was in a road cut north of Pimlico Pond. Similar till clumps were later noted at a few localities in the pitted plains of the Sagamore and Wareham quadrangles, west of the Cape Cod canal.

The individual masses of till range from a foot or two in length and thickness to more than 10 feet in longest dimension. They are generally more or less rounded or oval in shape although some have irregular or angular bounding surfaces. Commonly there are several clumps within a few yards of each other at any one locality, each entirely surrounded by sand and gravel (Pl. 2, fig. 2). Each locality is at the rim of a kettle hole. No till was found in the scores of gravel pits and road cuts on the smooth parts of the plain or in the many, small, sand pits near the cranberry bogs that have been developed in the furrows at a distance from the kettle holes. Evidently the till was formerly associated with the ice blocks responsible for the kettle holes near which it is now found.

The rock types represented by the pebbles and angular fragments embedded in the sand and clay of the till clumps are like those of the till in the Buzzards Bay moraine, both with regard to variety and relative abundance of the different components. Ordinarily the compactness and coherence of the till are such that the till clumps stand out on the faces of the artificial cuts after exposure to weathering for only a few days or weeks, during which time the surrounding sand and gravel is loosened and blows away or slumps toward the foot of the cut. The surface of each till clump is commonly stained to a depth of 2 to 8 inches with bright red iron oxides, and streaks of the same color penetrate nearly or quite to the center of some of the clumps. Moreover, the sand and gravel immediately adjacent to the till are ordinarily

oxidized similarly so that the same red color appears as an aureole surrounding the till at several localities.

All the clumps that have been observed are in the uppermost 6 to 10 feet of the gravel forming the plain, and some extend upward practically to the surface. Where overlain by sand and gravel the covering is ordinarily so thin that in contrast to the subjacent material the bedding has been obscured or destroyed by frost action, the penetration of roots, and the activities of animals. The filling between clumps is relatively loose, non-bedded, or irregularly bedded sand and fine gravel. The uppermost beds of gravel beneath the clumps are wrinkled into wavelike crests in the spaces 1 to 4 feet wide between the bases of adjacent clumps. At the Teaticket locality, the long till clumps stand on end. Near Bear Hollow Farm the long axes of several clumps are all inclined in the same direction. At the National Guard locality, where the till masses lie more or less horizontally, their ends are twisted into a scroll-like form. Where the cuts were deep enough to expose several feet of thickness of gravel below the till, the gravel layers appeared quite undisturbed at depths greater than 2 or 3 feet beneath the base of the clumps.

Evidently the till masses came into place very late in the building of the plain, and considerable pressure was exerted upon at least some of them as they assumed their present positions. Such pressure may have resulted from contraction and expansion of the ice near the top of a stranded ice block exposed to rapidly fluctuating air temperatures. Till dropped directly on the gravel floor beneath the edge of a melting block of ice may have thus been squeezed into place. Moreover, the streams of meltwater on the surrounding plain would occasionally undercut the sides of the dwindling remnants of the glacial ice and cause chunks of it to fall or slide downward and outward, thus shoving the till vigorously into the gravel to produce the wrinkles in the subjacent beds.

SUBAERIAL FURROWS

The surface of the Mashpee pitted plain is not only pock-marked by kettle holes, but it is also scored by long furrows that head about midway up the plain and extend to its outer edge. The longest measures about 10 miles. The lower parts of the larger furrows are now drained by small streams, such as Coonamessett River and Childs River, but these are not actively enlarging the furrows today. The lower end of each is flooded by the sea and forms a long, narrow bay, such as Great Pond, Green Pond, Bowens Pond, and Eel Pond (Fig. 3).

Erosion rather than deposition is responsible for these furrows, since they are actually cut into the sand and gravel layers of the plain, as indicated by the truncated edges of gravel beds seen in exposures on the

walls of furrows. Two of the best exposures available in 1939 were in road cuts half a mile southeast of Jenkins Pond (Pl. 2, fig. 3) and a quarter of a mile southeast of Round Pond. Furthermore, the furrows have the form of stream-cut valleys. Each heads in a narrow, shallow, V-shaped cut. Down valley, most of them increase to approximately 1000 feet in width and 25 feet in depth. Their floors are broad and smooth. Short, V-shaped tributaries enter the main valleys and give each system a stunted dendritic pattern like the veins of a narrow leaf.

The close spacing of the furrows and subparallel orientation of all their main courses are remarkable. Some adjacent furrows are so close where they enter Nantucket and Vineyard Sounds that the intervening flat-topped remnants are rarely as much as half a mile broad. All are oriented within 20° of south. Those near the western edge of the plain are nearly parallel to the Buzzards Bay moraine. Such an arrangement of valleys is not like the dendritic "crows-foot" pattern formed in the normal processes of stream erosion, and it, therefore, requires some special explanation.

As many furrows are not now occupied by streams, present-day erosion is totally inadequate to account for these forms. As the headwaters are not directly connected with areas of moraine, and as the pattern of the furrows is not braided and therefore is not indicative of overloaded streams, the furrows were not cut by the glacial melt water that constructed the plain. It is possible that, after the plain was essentially complete, the extra supply of water from ice melting west and north of the moraines saturated the loose gravels. This ground water might have reached the surface in springs part way down the fan and flowed on the surface from there to the sea. A second possibility is that there was a period of barren, treeless climate, caused either by drouth or cold, during which ephemeral rains cut shallow arroyos in the gravels. It is also possible that the condition of the ground, frozen while the ice sheet still covered the interior of New England, made all save surface gravels impermeable to water and induced shallow cutting in closely spaced channels.

PERSISTENCE OF BURIED ICE

The duration of all these events which contributed to the shaping of the present Mashpee pitted plain is very impressive. Following the isolation of large masses of ice from the wasting ice sheet, braided streams of melt water deposited sand and gravel around and over most of the outlying ice masses. The period required for the construction of such a large plain, which now contains more than 2 cubic miles of gravel, must have been many tens and perhaps hundreds of years.

The surface materials of the plain include many thousands of stones which are plainly cut, grooved, and faceted by the wind. These may be found singly or in quantity in nearly half of the pits now excavated. The deep etching of these ventifacts and their abundance impress all who see them with the vigor and duration of wind action that must have taken place. No comparable wind action is effective on the Cape today. Many of these ancient ventifacts are actually in the beds of gravel near the surface of the plain, but they could not have rolled far in the streams that built that portion of the plain or the delicate polish and edges made by the wind would have been obliterated. It is necessary, then, to assume that the deposition of the upper layers of the plain alone required many tens and perhaps hundreds of years during which wind action was very intense.

Buried ice persisted under the sand and gravel of the plain all the time that the wind was thus active, for none of the sand or silt that must have supplied abrasive tools for the sandblast fills the kettle holes. Furthermore, the furrows were cut before the ice entirely disappeared. Furrows are found to enter and exit from groups of kettle holes with no change of direction. There is no sign of drainage from one kettle hole to another. Dozens of small pits occur on the floors of furrows, and large pits may interrupt and separate a whole headward section of furrow from its lower flooded end at the sea. Had these low kettle-hole depressions existed when the furrows were cut, they would certainly have been filled with stream gravel to the level of the channel floor. As this is not true, the ice must have remained after the plain was built until the furrows had been cut.

Buried ice remained while the adjacent moraines were superimposed upon the edges of the plain. Thin active ice moved forward over this gravel and deposited the till of the Buzzards Bay and Sandwich moraines without destroying the layers of gravel beneath. Masses of stagnant ice caught under the first outwash gravel remained until the active ice overrode them, inasmuch as certain large kettle holes already described extend from the exposed surface of the fan into the moraine-covered area. The ice finally melted away after both the gravel of the fan and the till of the moraine had been deposited.

DROWNED SHORE LINE

The final event affecting the present aspect of the Mashpee pitted plain was drowning by the sea. At no time since completion of the plain has sea level been higher on this portion of Cape Cod than it is today, for no sand bars were found above the modern shore line. These would certainly be present had wave action once been higher on such

a sandy shore. Moreover, none of the kettle holes at the lower edge of the plain are filled by wave wash. On the other hand, submerged cedar stumps have been dredged from below low-tide level at Witchmere Harbor in Harwichport, 15 miles east of this pitted plain (Chute, 1939, p. 25), and at Centerville (Sawyer, 1932) near its eastern edge. The lower end of each furrow is obviously drowned; no process of shore erosion or construction could produce such a deeply indented shore line. Following the rise of the sea to its present stand with respect to this part of Cape Cod, sand bars were built across many of the bays and estuaries.

Low wave-cut cliffs facing narrow sea beaches have been eroded along the low margin of the plain. If the original gentle surface of the plain be projected beyond the present cliffs until it touches sea level, it appears that the plain once extended at least a quarter and perhaps half a mile farther seaward. Waves and shore currents have beaten the shore line back by this amount. At distances greater than half a mile off the Falmouth shore, or one mile off the shore at Cotuit, the Coast and Geodetic Survey charts (no. 1209 especially) show a submarine slope far too steep to be the original surface of the pitted plain. If the pitted plain formerly extended beyond that line, recent excavation by currents, or wave cutting at some lower stand of the sea, has modified the bottom contour too much to permit reconstruction of the original outer margin of the plain.

PART III. LATE GLACIAL VENTIFACTS

OCCURRENCE AND DISTRIBUTION

During these investigations on western Cape Cod thousands of wind-worn stones or ventifacts (Bryan, 1931) were found embedded in the various till masses and glacio-fluvial deposits. Reconnaissance at other widely scattered localities on or near the Cape—Orleans, Brewster, Wellfleet, Chatham, and Plymouth—disclosed a similar remarkable abundance of ventifacts in the glacial and postglacial deposits. This confirmed reports of several writers (Woodworth, 1894; Davis, 1895; Bryan, 1932) that aeolian action was prevalent in this part of New England during or since the retreat of the last Pleistocene ice sheet. More than 2000 of these wind-sculptured stones were collected and studied, not only because they exhibited great variety in shape and in type of abrasion but also for the information they might provide concerning the age of the deposits enclosing them.

The ventifacts were found in various relations to their surroundings. Their mode of occurrence may be described under four categories.

(1) Some were lying on the surface. Many of the stones thus noted

appear to have reached their present position largely through overturn of the upper few feet of material by frost action, though some may represent residual accumulations from slope wash. Ventifacts are also fairly common on beaches derived from shoreline erosion of till or gravel bluffs, as at Nobska Point (Woods Hole quadrangle), Wings Neck (Onset quadrangle), Falmouth Heights (Falmouth quadrangle), and Sagamore Heights (Sagamore quadrangle).

(2) Others occur in the upper soil zone and in the disturbed and weathered zone, or "warp." The warp is a thin mantle over undisturbed till or over gravel layers. Where the thickness of the disturbed material was observed, it generally approximated 2½ to 3 feet and rarely was as great as 4 feet. There is good reason to believe, moreover, that very little material has been removed by slope wash or wind action at many localities. The depth of frost action was not very great, therefore, despite the periglacial climate that prevailed during recession of the ice.

(3) Some of the ventifacts are at the top of the glacio-fluvial gravels and beneath well-sorted sand, several feet thick. Although this mode of occurrence is uncommon, it is well exemplified in a gravel pit overlooking Whitcomb's Bog (Pocasset quadrangle) (Pl. 3, fig. 1). There, scores of ventifacts of many sizes, including small boulders, lie closely packed along a single horizon, apparently undisturbed since they were cut. Only the top and sides of each stone display signs of wind work.

(4) Many ventifacts are embedded in undisturbed till and gravel. Specimens were collected from layers of well-bedded gravel, 2 to 8 feet below the disturbed zone at Falmouth Heights, 10 feet below at Whitcomb's Bog, 8 to 20 feet below at Nameloc Heights (Sagamore quadrangle), and from till, 2 to 8 feet below the "warp" at Woods Hole, at Gunning Point (Woods Hole quadrangle) and at Wings Neck. The ventifacts embedded in till have, of course, a random orientation, and many show signs of abrasion subsequent to wind-cutting. Presumably, they were picked up during minor readvance of the ice which deposited the till that enclosed them. Some of the stones embedded in gravel are scattered at random and with varied orientations of the wind-smoothed surfaces. Many of them show varying degrees of abrasion subsequent to windcutting, which must in part at least be a function of the distance they were carried by the melt waters that brought them to their present position. Other ventifacts embedded in gravel occur in groups along certain horizons and appear to have been simply buried with little if any disturbance by the superjacent layers of gravel or sand.

Wind-cut stones can be found associated with glacial and glacio-fluvial deposits almost everywhere on Cape Cod and in adjacent areas of New England (Shaler, 1889, Pl. X; Woodworth, 1894), yet they are

FIGURE 1. GRANITE GNEISS BOULDER WITH FLUTING TRANSECTING FOLIATION
Boulder is at top of bedded gravel beneath about 6 feet of well-sorted sand, exposed in pit at side of Whitcomb's Bog.

FIGURE 2. GRANITE BLOCK ON SHORE OF CAPE COD BAY
Near Stage Point, 3 miles north of Sagamore quadrangle. Pits and grooves radiate from windward end in foreground. They are 1½ inches deep where impact of wind-driven sand was direct and grade to shallow flutings on sides of stone.

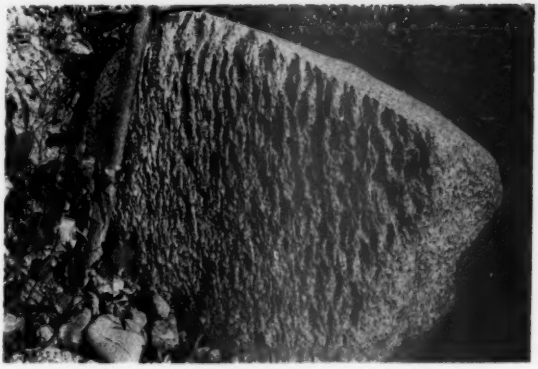


FIGURE 3. DEEPLY SCULPTURED GRANITE BOULDER
Grooves and pits are 1 to 2 inches deep.

VENTIFACTS PHOTOGRAPHED IN THE FIELD

FIGURE 1. VENTIFACTS TYPICAL OF THOSE DISPLAYING PYRAMIDAL FORM
Quartzite at left rear has 16 interfacet edges, only a few of which are shown. Small ventifacts in the foreground are aplites; the one at right rear is a felsite porphyry.

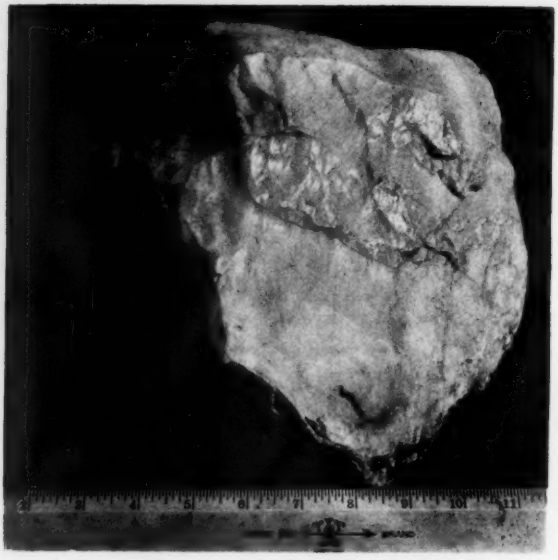


FIGURE 2. HIGHLY POLISHED SURFACE OF QUARTZITE VENTIFACT
Light reflected from conchoidal surfaces below sharp interfacet edge near top shows patchy preservation of cellophanelike polish.

PYRAMIDAL VENTIFACTS AND WIND POLISH

very unevenly distributed. One pit may yield hundreds, whereas pits near-by seem totally devoid of them. Even within a single exposure there is no uniformity in occurrence. Ordinarily they are sparsely distributed in both till and gravel, but patches of till a few feet in diameter and certain parts of single gravel layers may yield hundreds of specimens. In general, the ventifacts seem to be more common in the upper 4 to 8 feet of the deposits.

ROCK TYPES REPRESENTED

The glacial deposits of Cape Cod were derived from highly varied lithologic units comprising the bedrock in parts of eastern and northern New England. It was not surprising, therefore, to find more than 30 distinct rock types among the wind-worn stones. The rock species which were separately recognized in the writers' collection are listed below; opinions, based on lithologic comparisons, concerning the probable derivation of some of them are indicated. Such opinions are shared with several geologists conversant with the areal geology of New England who have visited these areas or examined specimens from them.

The lithologic types distinguished among Cape Cod ventifacts are:

Vein quartz (pure milky, blue brecciated, chloritic, jasperized, and colorless varieties)
Pegmatite and graphic granite (pink, yellow, and gray varieties)
Quartz-pink-feldspar gneiss
Aplite
Fine-grained granite and granite gneiss (probably fine-grained facies of the coarser types)
Medium to coarse granite and granite porphyry (massive to gneissose)
    Dedham granodiorite type
    Quincy granite type
    Types common in Maine ?
    Milford granite type ?
Diorite and gabbro-diorite
    Salem type
Diabase
    Medford type
Fine-grained mafic rocks (massive, porphyritic, and amygdaloidal)
Volcanic rocks of rhyolitic to intermediate composition
    Agglomerates and breccias
        Types found at Nantasket
    Felsite porphyries
      and
    Flow rhyolites
        Lynn volcanic types and type associated with Quincy granite
    Amygdaloidal
Conglomerates
    Roxbury conglomerate types
Quartzites (white, brown, blue-black, red, black and white varieties)

Argillite, phyllite, and slate (gray, green, red, laminated varieties)
Cambridge slate types
Biotite gneisses
Schists (biotite, chlorite, black and white quartzose laminated)

SIZES AND SHAPES

The wind-cut stones range from pebbles less than an inch in diameter to enormous boulders. Many large granite blocks have more than 20 square feet of wind-etched, fluted, or polished surface. Excellent examples may be seen on the beach at Wings Neck, and at Sagamore Heights. Many others were found in place in glacial till or gravel. The largest single wind-cut surface seen is the top of a huge boulder, known as Sacrifice Rock, in Bournedale, which measures approximately 30 by 20 feet. Its entire upper surface was polished and slightly fluted by wind abrasion.

So great a variety of shapes was found in the collection as a whole and among the stones gathered at single localities that there is no basis for conclusions concerning constancy or variability of winds, such as have been developed by other writers from experimental and field evidence. Eddying currents must have been produced at many places in the wind stream by the larger stones, and smaller ones lying on their leeward sides were thus protected to varying degrees. The shapes produced on these would consequently be far different from those which result when single, isolated specimens are attacked by sand blown across a desert floor or a laboratory table. Furthermore, the study of Cape Cod ventifacts revealed nothing to substantiate Heim's conclusion (1922, p. 349) that the number of facets that will develop on a stone undergoing wind abrasion is determined by the shape of its base. On the other hand it appeared that, in this collection at least, the shape of the original fragment was the most important factor controlling the present shape of the ventifact.

The following general shapes were found among the wind-faceted stones of the Cape, along with many forms intermediate between or representing combinations of these types:

(1) *Einkanter* (single facets bevelling rounded stones)—single, flat or convex facets truncating the curvature of more or less symmetrical, water-worn cobbles. These, however, are uncommon; most of the stones show truncation on more than one side.

(2) *Firstkanter* (ridge-shaped)—two major facets sloping away from each other and forming a ridge; includes shapes resembling spearpoints or arrowheads (Pl. 7, fig. 2).

(3) Pyramidal—three or more facets sloping from a common center (Pl. 4, fig. 1); includes *dreikanter*, *fünfkanter*, and so on. Contrary to

expectations from the impression generally conveyed in textbooks, *dreikanter* are the exception among these wind-cut stones. Less than 20 were found, although many others approach that shape.

(4) Triquetous or brazil-nut—three facets, commonly slightly curved, intersecting at about 60° (Pl. 5, fig. 1).

(5) Polygonal—number of facets variable and probably determined largely by the number of fracture surfaces on the block prior to sandblasting.

(6) Irregular—shape and number of facets very variable; this group may include early stages in the formation of the types listed above.

The length of time different facets on the same specimen were exposed to wind erosion, as judged by the depth of pits and flutings and the amount of material removed, varied considerably. This is more readily explained as a result of rolling of the specimen than as an effect of variable winds.

Even less of a tendency to uniformity in shape was discernible among the smaller stones, though they have been subjected to a more varied attack such as would affect the larger ones in time through continued erosion and rotation. This opens to question the concept advanced by some writers (Schoewe, 1932, p. 125) that the brazil-nut or triquetous shape represents the ultimate form that all ventifacts, regardless of their texture, should acquire in time, given a strong wind of constant direction.

EVIDENCE OF SANDBLASTING

The evidence given below makes inescapable the conclusion that the stones considered in this paper are truly ventifacts, shaped and scoured to varying degrees by wind-borne sand and dust. The features listed have all been ascribed by various writers to aeolian abrasion; they have been observed on ventifacts from modern deserts and beaches; and they are difficult to explain except as results of sandblasting. Consequently, they serve as useful criteria for recognizing ventifacts. All the specimens collected by the authors show one or more of them.

(1) Polished surfaces. Very few of the specimens have the high luster characteristic of ventifacts from modern deserts, but the preservation of patches of such a polish on some (Pl. 4, fig. 2) and of a somewhat duller luster on at least parts of the rest suggests that originally most of the wind-cut surfaces were highly polished. Gentle buffing with a soft cloth may be necessary to remove a film of dust that conceals the gloss. Commonly the polish is preserved only by quartz grains in granitic rocks; but some feldspar grains also have a somewhat duller luster. Quartz and quartzite are more lustrous in general than the feldspathic rocks.

(2) Smooth, greasy surfaces. Except where this characteristic has been destroyed by weathering or abrasion, wind-carved surfaces are smooth and feel distinctly greasy, like talcum powder. This is true not only of the larger faceted and cuspate surfaces but also of the inner parts of grooves, pits, and hollows, and along the windward sides of low projections produced by differential etching.

(3) Wind-smoothed faces and wind-cut facets. These range from less than half an inch square to many square feet. Most are either concave or convex, but a few are essentially flat. The number of such surfaces on any single stone varies considerably. Some of the facets probably represent merely modified fracture surfaces, but those that bevel rounded cobbles and those that form pyramidal shapes were clearly produced by wind abrasion (Pl. 3, fig. 3; Pl. 4, fig. 1).

(4) Sharp interfacet edges. These range from knifelike sharpness on some quartzite cobbles to rounded and pitted ridges on water-worn specimens (Pl. 4, fig. 1; Pl. 5, fig. 2). Edges formed at the junction of fluted facets are serrate (Pl. 6, fig. 1).

(5) Differential etching. Veins, phenocrysts, inclusions, fragments in volcanic rocks, quartz blebs, laminae, pebbles in conglomerate, and amygdules, commonly stand in relief on windswept surfaces or are etched below the general level of their surrounding media (Pl. 7, fig. 1). In either case a well-defined luster, a smooth and greasy feel, and occasional small-scale flutings show that they were produced by differential etching in a sandblast rather than by differential solution during weathering. Some differential solution may have preceded the wind action on a few of the stones so that the irregularities were accentuated by the wind, but aeolian abrasion alone could have produced their present shapes. Several ventifacts with such etching have bizarre shapes, representing small-scale examples of the peculiar erosion forms seen on landscapes in arid and semi-arid regions (Pl. 6, fig. 2).

(6) Fluted surfaces. Differential resistance of the various constituent minerals in granular igneous rocks expresses itself in groups of discontinuous, subparallel, concave grooves or flutings. These form more or less denticulated surfaces that range from almost microscopic indentations to 1½ inches in depth (Pl. 3, figs. 2, 3; Pl. 6, fig. 1). They were carved by wind moving essentially parallel to the rock surface on which they occur. Their orientation is independent, therefore, of pre-existing structure in the rock and is parallel to the wind motion. They may cut diagonally across gneissic banding, but on some specimens, or on parts of one specimen, they may fortuitously follow it.

On the coarser-textured ventifacts, fluting is universally present. It may be a subordinate feature of well-formed, wind-carved facets (Pl. 5,

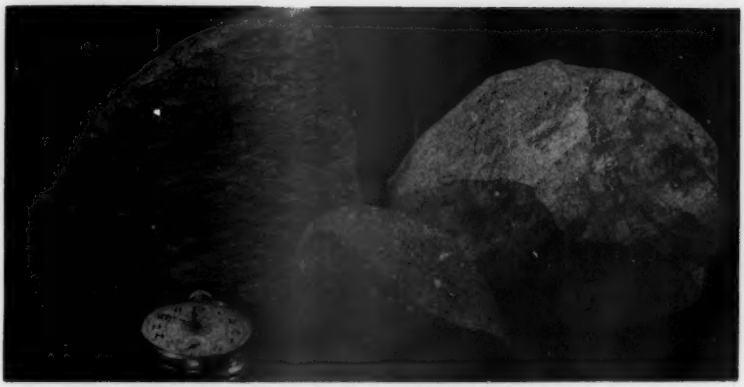


FIGURE 1. VENTIFACTS SCULPTURED FROM VOLCANIC ROCKS
Shallow flutings on basalt porphyry at left were controlled by etching out of phenocrysts. Recrystallized rhyolite tuff at right is denser, and fragments were as resistant as matrix. Phenocrysts in basalt porphyry in middle exerted little influence on ultimate triquetous form.

FIGURE 2. VENTIFACT CARVED PARTLY FROM GRANITE AND PARTLY FROM DIKE ROCK
Sharp edges and smooth facets are typical of fine-textured rocks, irregular pits and flutings of coarse texture.

INFLUENCE OF ROCK TEXTURE ON FORMATION OF VENTIFACTS

FIGURE 1. DEEPLY FLUTED GRANITE GNEISS
Each side is a fluted facet; orientation of grooves indicates rotation of stone on long axis.

FIGURE 2. DIFFERENTIAL ETCHING OF THREE LITHOLOGIC TYPES
Greenstone at left, porphyry in middle, and fine-grained gneiss at right. Trails on leeward side of projecting phenocrysts are notable on surface of porphyry.

GROOVED AND ETCHED VENTIFACTS

fig. 1) or it may occur on rather flat surfaces which represent modified fractures. Fine-textured rocks, such as volcanic porphyries or gneisses, containing elements of different hardness, may have fluted facets; but homogeneous, hard quartzite, aplite, and vein-quartz pebbles do not ordinarily show fluting.

The arrangement and character of the flutings may furnish excellent evidence concerning the wind direction and other conditions under which a certain ventifact was produced. For example, on some specimens the fluting is diagonal to interfacet edges (Pl. 3, fig. 2). This condition eliminates the possibility that such ventifacts originated when opposing facets were developed simultaneously by a sandblast directed essentially parallel to the interfacet edge. It also renders inapplicable to these particular stones the observation of Woodworth (1894, p. 69) that sloping facets may be produced in a position more or less normal to the wind. Again, the orientation of flutings on adjacent facets may show not only that a stone was dislodged and rotated into a new position before each succeeding facet was produced, but it also may indicate the directions and axes of such rotation (Pl. 6, fig. 1).

(7) Pits and cuspate hollows. Shallow to deep pits may occur singly or in groups on both smooth and fluted facets, representing the etching out of phenocrysts or other less resistant portions of the stone. The size, shape, and distribution of the hollows are controlled by patterns of nonresistant elements in the rock (Pl. 5, fig. 1).

Several fluted and faceted granite boulders show that they split the wind. On their windward ends, where the surface of impact was nearly normal to the sandblast, deep pits were formed by etching out of feldspars. The pits deepen to leeward, and some have overhanging, roof-like projections of more resistant quartz or feldspar at their leeward terminations. On surfaces curving upward to facets on the top and sides of the boulders (where the sandblast moved more nearly parallel to the erosion surface) these pits become progressively shallower and increasingly elongated parallel to the wind until finally they grade into ordinary flutings on the flatter parts of the facets. The pits, hollows, and flutings display a definite tendency to radiate from the windward ends of the boulders (Pl. 3, fig. 2).

One specimen of dense greenstone has peculiar cuspate hollows on its windward side directly beneath a projecting ridge of quartzose vein material that forms its upper surface. These cusps also deepen to leeward more rapidly than those described above so that they are essentially inclined and cup-shaped. Lateral coalescence of several of them formed a scalloped edge in front of the quartz overhang. Smoothness and polish of the bottoms of the cusps, of the windward face of the

rock, and of the projecting ledge show that this phenomenon is properly to be attributed to aeolian action rather than to solution.

Quartz and quartzite ventifacts commonly have broad, shallow cusps on well-developed facets. These are not of the same origin as the pits described above but were probably produced by modification of the curvatures normally formed by conchoidal fracture.

(8) Effects on small stones. Almost all the characteristics of wind abrasion found on the larger ventifacts are duplicated on those which range from 2 inches to less than half an inch in diameter (Pl. 7). These smaller ventifacts show likewise that their texture and mineral content were important in controlling the development of fluting, pits, and facets and that rotation during wind-cutting determined the number and perfection of their facets. These small-scale "replicas" were, however, affected on all sides more commonly than their larger counterparts, doubtless because they became dislodged and were rotated more readily and more frequently.

ABRASION DURING SUBSEQUENT TRANSPORTATION

Many of the ventifacts were modified to varying degrees by abrasion during glacial or glacio-fluvial transportation. This is seen in destruction of the polish, rounding of the interfacet edges, lowering of the ridges between adjacent grooves or flutings, and replacement of the smooth, greasy surfaces by lusterless, minutely pitted ones that have a harsh and gritty feel. Several stones were found whose only remaining evidence of wind-cutting consisted of patches of unmistakable fluting, glossy and smooth in their deeper portions. Hundreds of ventifacts, on which facets and other indications of wind-scouring were almost obliterated by weathering and abrasion, were seen but not collected. Still other hundreds were no doubt overlooked because they resemble too closely the normal rounded to subangular stones of the glacial deposits. This is probably true also in other glaciated or driftless areas where the existence of ventifacts associated with the Pleistocene deposits has not been reported. There is little evidence of abrasion of this type on the abundant ventifacts in the upper 8 to 10 feet of the gravel deposits. They could not have been carried far, therefore, after the wind had shaped them. The lack of glacial striae on the modified facets of ventifacts embedded in till is a further indication of short transport from the places where sandblasting occurred to present positions in the till.

WIND-CUT SURFACES TRANSECTING WEATHERED RIMS

General statement.—On a few ventifacts of differing lithologic types (one granite, two greenstone volcanics, one quartzite) the facets were developed across narrow limonitized and hydrated rims. These stones

FIGURE 1. DIFFERENTIAL ETCHING BY SANDBLAST
Rocks are greenstones and volcanic breccias.

FIGURE 2. SPEAR-SHAPED VENTIFACTS
Quartzite in middle rear has cusps on modified surfaces of conchoidal fracture; other specimens are granites and felsites.

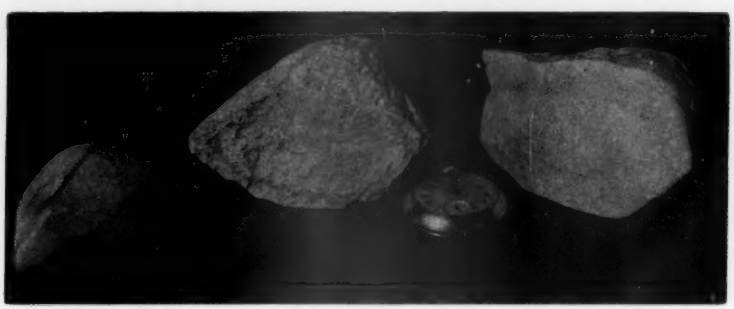


FIGURE 3. WIND-CUT FACETS TRUNCATING WEATHERED RIMS
Limonitized selvage on felsite porphyry in middle is cut by sandblasted surface at upper right; lower front face is fracture broken to show complete weathered rim. Other specimens are greenstones with wind-cut facets facing camera; darker zones are the result of weathering before exposure to sandblast.

VENTIFACTS OF VARIOUS SHAPES
Some showing weathering before sand-blasting.

were broken through with a hammer to prove that the rims were truly formed by centripetal penetration of weathering agencies rather than by superficial staining while the specimens lay embedded in the soil. The facets were cut deeply enough to expose the fresher cores of these rocks so that now the brown rims form a border selvage on the facets (Pl. 7, fig. 3). On one specimen the formation of the facet removed a whole side of the rounded stone so that no such border remains on that side. Another greenstone cobble was rolled and so deeply faceted on all sides that only a few patches of its weathered rim have survived.

The facets on most of the ventifacts do not show similar staining or comparable weathering. In fact, some have been so little affected by weathering since they were wind-cut that they have retained a greasy, lustrous surface. In a rock as susceptible to weathering as greenstone this can only mean that little time has elapsed since the facets were carved, unless an appeal be made to extraordinarily perfect protection from weathering processes.

Date of formation of weathered rims.—No certain way of dating these uncommon rims has yet been found. Since the weathered rims obviously were formed prior to the facets that cross them, and the facets were carved during recession of glaciers believed to represent the Wisconsin stage on Cape Cod, the weathering must be assigned to a pre-Wisconsin substage of glacial time. It does not seem likely that cobbles weathered during pre-Illinoian time could have survived Illinoian transportation, Sangamon weathering, and Wisconsin transportation. It is possible, however, that the present weathered rims are merely the still fairly compact inner parts of much thicker shells of decomposition which were largely removed during these stages. Too few similarly weathered rims were seen on boulders and cobbles that escaped wind-cutting to support the suggestion that the rims were formed during the short interval just preceding the wind-cutting—that is, between the time the stones were washed out of receding ice and the beginning of wind wear. Determination of the precise age of this weathering is a subject for further investigation. Nevertheless, the finding of such partly weathered stones in glacial drift among the myriads of almost fresh ones increases the necessity for taking into account the possibility of preglacial or interglacial weathering of some stones in applying the comparative weathering of constituent stones as a criterion of age in separating drift deposits (MacClintock, 1940).

SIGNIFICANCE OF VENTIFACTS IN DATING DEPOSITS

The widespread occurrence of ventifacts within the till masses and outwash gravels of western Cape Cod indicates that they were sculptured

under periglacial conditions during recession and local readvance of the ice. Modifications of the wind-scoured surfaces during transport after wind-cutting could have developed only if aeolian abrasion was in progress during construction of each of the successive glacio-fluvial landforms and prior to the development of each moraine in which ventifacts are embedded. Stones lying on a surface of outwash were sand-blasted and then picked up in a readvance of the ice to become incorporated in the till of the resulting push moraine. This sequence of processes was doubtless repeated many times during the recession of the ice from its terminal position.

The ventifacts support and corroborate other lines of evidence that no great length of time has elapsed since the deposits of western Cape Cod were made. The freshness of all the wind-cut surfaces, including those across earlier weathered rims, and of the associated stones, argues strongly against a date earlier than the Wisconsin stage for their formation. The preservation of polish on many ventifacts in the surface zones of the deposits points to the same conclusion. These stones in the disturbed "warp" cannot be assigned to post-Wisconsin wind action because there is so little wind-drifted sand on the surface, even in the bottoms of numerous kettles—no anchored dunes, loess deposits, or patches of aeolian sand capable of forming the many thousands of ventifacts. Stetson (personal communication, 1940) reports abundant frosted and well-sorted sand (probably aeolian) on the bottom of Vineyard Sound, which may represent the missing abrasive. If so, it blew off the present area of the part of the Cape under consideration before the kettle holes were formed, for such hollows could not have escaped at least partial filling had they existed when the ventifact-producing sands swept across this region. These stones must have been shaped, therefore, before all the buried blocks of ice had disappeared. This does not harmonize with the suggestion advanced by some geologists that the ventifacts were formed at the surface in postglacial time and churned back with random orientations into the ground by vigorous frost action. The evidence given above seems to indicate, on the contrary, that both the ventifacts and the deposits that contain them belong to a late Wisconsin stage.

CLIMATIC IMPLICATIONS

Although in abundance and perfection the ventifacts are clear indicators of strong winds blowing down from oscillating margins of late Wisconsin ice, they do not signify an arid climate. The thickness, areal extent, coarseness, and channeled character of outwash gravels on western Cape Cod indicate strong discharge of meltwaters, periodically at least, during the accumulation of such debris. Nevertheless, ablation

must also have been an important process in the downwasting and disappearance of the ice because there was apparently little subaerial erosion of pitted outwash plains by meltwaters while recessional moraines north of them were still being formed by active ice. The climate must have been somewhat arid, however, to permit large quantities of sand to be drifted about the area and to prevent the flourishing of vegetation which would have anchored the sand and would probably have caused more pronounced chemical decomposition of the constituents of the deposits than has been observed. That there was only slight to moderate abrasion of many of the ventifacts during glacio-fluvial transport strongly suggests short transfer, such as would occur in sudden flushing of meltwaters, torrential rains, or rapid shifting of channels. The character of subaerial valleys (furrows) on the pitted plains is strikingly similar to that of arroyos carved under semi-arid conditions. The authors conceive, then, of the sculpturing of these ventifacts and disappearance of the ice in a disagreeable, semi-arid climate characterized by sporadic heavy precipitation, strong winds, and low temperatures.

SUMMARY

Study of more than 2000 ventifacts embedded in undisturbed deposits and in disturbed surficial zones of western Cape Cod yields corroborative evidence that the deposits were formed during late Wisconsin time. The climate during recession of the ice and sculpturing of the ventifacts was cold and semi-arid. The present shapes of the wind-cut stones were determined by the shapes of the original fragments, textural and structural features of the individual rock, amount of dislodgment and rotation during sandblasting, duration of abrasion while a stone lay in one position, and degree of protection from direct impact afforded by adjacent stones. Because such a large number of lithologic types is represented among these ventifacts a variety of erosional effects which may be used as criteria of sandblasting were observed. Modification of wind-scoured surfaces during later glacial and glacio-fluvial transportation shows that aeolian abrasion occurred during construction of the various landforms made of outwash and prior to the development of each successive recessional moraine. Lack of an appreciable veneer of wind-drifted sand over the surface, even in the bottoms of kettles, indicates that cutting of the ventifacts was completed before all of the residual buried ice blocks had disappeared.

REFERENCES CITED

Bryan, Kirk (1931) *Wind-worn stones or ventifacts—a discussion and bibliography*, Nat. Res. Council, Rept. Comm. Sedimentation, 1929-1930, p. 29-50.

——— (1932) *New criteria applied to the glacial geology of southeastern Massachusetts* (abstract), Geol. Soc. Am., Bull., vol. 43, p. 176.

Chamberlin, T. C. (1883) *Preliminary paper on the terminal moraine of the second glacial epoch,* U. S. Geol. Survey, 3d Ann. Rept., p. 377-379, pl. 33.

Chute, N. E. (1939) *Geology of the coastline between Point Gammon and Monomoy Point, Cape Cod, Massachusetts,* Commonwealth Mass., Dept. Public Works, Spec. Paper 1, p. 1-26.

Davis, W. M. (1895) *Faceted pebbles on Cape Cod, Massachusetts,* Boston Soc. Nat. Hist., Pr., vol. 26, p. 166-175.

Heim, Arnold (1922) *Känguruh-Polituren in Australien,* Zürich, Société Géologique Suisse, vol. 17, p. 349-351.

Hitchcock, Edward (1833) *Report on the geology, mineralogy, botany, and zoology of Massachusetts,* p. 143-144.

Howe, O. H. (1936) *The Hingham red felsite boulder train,* Science, vol. 84, p. 394-396.

Johnson, Douglas (1925) *The New England-Acadian shoreline,* New York, p. 105-117, 295.

MacClintock, Paul (1940) *Weathering of the Jerseyan Till,* Geol. Soc. Am., Bull., vol. 51, p. 103-116.

Mather, K. F., Goldthwait, R. P., and Thiesmeyer, L. R. (1940) *Preliminary report on the geology of western Cape Cod, Massachusetts,* Mass. Dept. Public Works, U. S. Geol. Survey, Coop. Project, Bull. 2, p. 40-43, 46-52.

Sawyer, G. C. (1932) *A drowned forest of white cedar at Centerville, Cape Cod, Massachusetts,* mss. report to Dept. Geology, Dartmouth College, Hanover, N. H.

Sayles, R. W. (1939) *Upper till, two boulder clays, and interglacial flora on Cape Cod* (abstract), Geol. Soc. Am., Bull., vol. 50, p. 1931-2.

Schoewe, W. H. (1932) *Experiments on the formation of wind-faceted pebbles,* Am. Jour. Sci., 5th ser., vol. 24, p. 111-134.

Shaler, N. S. (1889) *The geology of Nantucket,* U. S. Geol. Survey, Bull. 53, p. 21-26; pl. X; p. 46.

——— (1898) *Geology of the Cape Cod district,* U. S. Geol. Survey, 18 Ann. Rept., pt. 2, p. 297-593.

——— and Davis, W. M. (1881) *Illustrations of the earth's surface, Glaciers,* Boston, p. 57-58.

Upham, Warren (1879) *Terminal moraines of North American ice sheet,* Am. Jour. Sci., 3d ser., vol. 18, p. 81-92, 197-209.

Wilson, J. H. (1906) *The glacial history of Nantucket and Cape Cod,* Columbia Univ. Press, p. 4, 7, 48-50.

Woodworth, J. B. (1894) *Post-glacial eolian action in southern New England,* Am. Jour. Sci., 3d ser., vol. 47, p. 63-71.

——— and Wigglesworth, Edward (1934) *Geography and geology of the region including Cape Cod, the Elizabeth Islands, Nantucket, Martha's Vineyard, No Mans Land and Block Island,* Harvard Coll., Mus. Comp. Zool., Mem., Cambridge, Mass., vol. 52, p. 1-322.

GEOLOGICAL MUSEUM, HARVARD UNIVERSITY, CAMBRIDGE, MASS.; BROWN UNIVERSITY, PROVIDENCE, RHODE ISLAND; ILLINOIS INSTITUTE OF TECHNOLOGY, CHICAGO, ILL.

MANUSCRIPT RECEIVED BY THE SECRETARY OF THE SOCIETY, AUGUST 13, 1941.

PRESENTED BEFORE THE GEOLOGICAL SOCIETY, DECEMBER 30, 1939.

PUBLISHED BY PERMISSION OF THE DIRECTOR OF THE UNITED STATES GEOLOGICAL SURVEY AND THE COMMISSIONER OF PUBLIC WORKS, COMMONWEALTH OF MASSACHUSETTS.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 53, PP. 1175-1186, 2 FIGS. AUGUST 1, 1942

ORIGIN AND STRUCTURE OF THE LIVINGSTON IGNEOUS ROCKS, MONTANA

BY WILLARD H. PARSONS

CONTENTS

	Page
Abstract	1175
Introduction	1176
Purpose and location	1176
Acknowledgments	1176
Livingston formation	1177
Pyroclastic rocks	1177
Types and extent	1177
Composition	1178
Age	1178
Origin of extrusive series	1179
General statement	1179
Volcanic vents	1179
Explosion breccias	1179
Mud flows and stream deposits	1181
Intrusive rocks	1181
General considerations	1181
Iron Mountain laccolith	1181
General statement	1181
Contact metamorphism	1181
Hydrothermal veins	1182
Dikes and pipes	1182
Structure	1182
Relation to older rocks	1182
Relation of younger rocks	1183
Structures within the volcanic series	1183
Summary of conclusions	1184
General considerations	1184
Summary of events	1184
References cited	1185

ILLUSTRATIONS

Figure	Page
1. Sketch map of the Beartooth Mountain area	1176
2. Geologic map and cross sections of the Deer Creek-McLeod area, Montana	1180

ABSTRACT

Late Cretaceous andesitic volcanic rocks immediately north of the Beartooth Mountain uplift and included in the Livingston formation originated from many small vents near the headwaters of the Deer creeks. The series is composed of volcanic

breccias and tuffs near the vents and of volcanic conglomerates and sandstones farther away. The latter are interbedded with Judith River to Lance (upper Cretaceous) sediments. Pipes, dikes, and laccoliths of dacite, andesite, and diorite cut the pyroclastics and near-by sediments. Some breccia beds and the underlying sedimentary rocks have been tilted by the intrusion of several laccoliths prior to the Laramide orogeny. Other pyroclastics are later than the laccoliths but pre-Lance. Igneous activity had ceased before the main Beartooth thrusting.

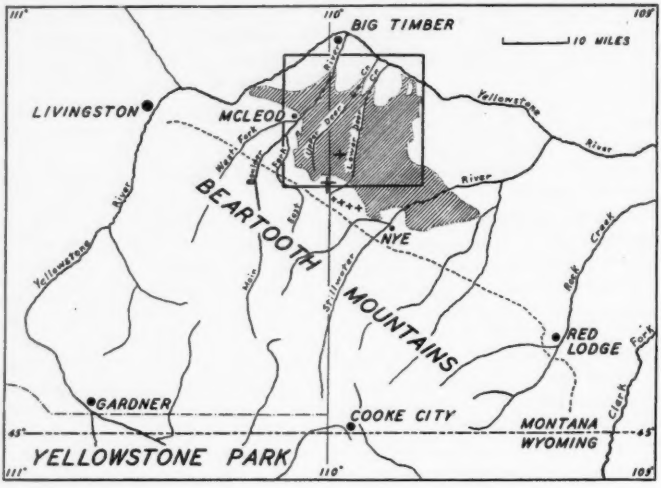


FIGURE 1.—*Sketch map of the Beartooth Mountain area*

Showing location of the Livingston igneous series (diagonal lined area). Broken line is approximate position of the Beartooth Mountain front. Crosses represent laccoliths (larger crosses = larger laccoliths). Small rectangle demarcates area of Figure 2, and area discussed in this paper.

INTRODUCTION

PURPOSE AND LOCATION

This study deals with the structure, origin, and age relations of the pyroclastic and intrusive rocks, the former a part of the Livingston formation, in the Deer Creek-McLeod area, Montana. The location of this area is shown in the accompanying sketch map (Fig. 1). These rocks represent one unit of the many late Cretaceous volcanic rocks of western Montana.

ACKNOWLEDGMENTS

Field work was begun in the summer of 1937 as part of a project under a grant from the Penrose Bequest of The Geological Society of America to Dr. M. H. Stow. Financial assistance from the Yellowstone-Bighorn

Research Association made possible field work in July 1938. The field work was completed during the summer of 1939 with the aid of a grant from the Penrose Bequest of The Geological Society of America for which the writer is duly grateful. Thanks are due Professor W. T. Thom, Jr., who suggested the problem, and Dr. Erling Dorf, who identified several plant fossils. The writer is especially indebted to Dr. Marcellus H. Stow for numerous profitable discussions and suggestions concerning the field work. George Phair was a capable field assistant in 1939 and did part of the petrographic work on the Iron Mountain intrusive as an undergraduate thesis at Hamilton College. The writer also wishes to thank Henry and Jack Eagle, Joe Kerr, Robert Andrews, and John Weitz for their very able field assistance.

LIVINGSTON FORMATION

The name Livingston formation was first applied by Weed (1893) to about 7000 feet of sandstones, shales, and conglomerates north and east of Livingston, Montana. Later studies by Stone and Calvert (1910) and others have shown that this formation grades laterally into the Claggett, Judith River, Bearpaw, Lennep, Lance, and Fort Union formations and has resulted in considerable remapping of Livingston formation areas. Present-day use of the term Livingston formation (Vhay, 1939) often refers only to Weed's (1893, p. 26-29) agglomerate member of the Livingston,—a member mapped as basic andesitic breccia in the Livingston Folio (Iddings and Weed, 1894) and including all the pyroclastic rocks discussed in this paper. Considering these changes in the use of the name Livingston, the writer has called these pyroclastics the Livingston igneous series to avoid further confusion.

PYROCLASTIC ROCKS

TYPES AND EXTENT

The Livingston igneous series comprises breccias, tuff-breccias, lapilli-tuffs, tuffs, vent agglomerates, tuffaceous sandstones, and volcanic conglomerates. Throughout this paper the writer uses the classification and terminology of pyroclastic rocks suggested by Wentworth and Williams (1932, p. 45-51). The areal extent of this series is shown on Figure 1. This pyroclastic outcrop corresponds approximately to the agglomerate area on the map of Stone and Calvert (1910, Pl. 7, facing p. 555).

Breccias, occasional tuffs, and vent agglomerates make up most of the series in the area at the headwaters of Upper and Lower Deer creeks and Bridger Creek (Fig. 2). These breccias show essentially no bedding. The fragments vary greatly in different beds but average 2 to 3 inches in diameter with a maximum of 3 feet locally. The matrix of the breccias

is usually a crystal tuff. Tuffs and lapilli-tuffs are not numerous. These pyroclastics are cut by breccia-filled plugs, and vent agglomerates.

Eastward along the Stillwater River near Nye (Fig. 1) the series contains well-bedded breccias, tuffs, and intercalated tuffaceous sandstones (Vhay, 1939, p. 434). Northward toward the Yellowstone River and westward along the Boulder River well-bedded volcanic conglomerates and tuffaceous sandstones predominate. Due to their lenticular character single beds do not extend far. Cross-bedding is not uncommon in the tuffaceous sandstones.

This pyroclastic series is at least 2000 feet thick in the Deer Creek headwaters region and thins to the north and west to approximately 50 feet at the Yellowstone River.

COMPOSITION

The breccias, tuffs, and volcanic conglomerates are all andesitic. Coarse and fine fragments alike consist of augite andesite, hornblende andesite and less commonly dacite and basalt. The andesites are composed of augite, hornblende, labradorite, andesine, minor biotite, magnetite, apatite, and orthoclase, with augite, hornblende, and labradorite as phenocrysts. The matrix is composed of fragments of these same minerals. Most of the pyroclastics contain fragments of all these types suggesting admixing from various contemporary vents.

The Enos Mountain breccias are exceptional, for they consist almost entirely of biotite dacite and have occasional limestone fragments. The dacite is composed of biotite, hornblende, andesine, orthoclase, quartz, magnetite, and apatite, the first three occurring as phenocrysts.

A few thin basalt and andesite flows are locally interbedded with breccias near vents. Undoubtedly more exist than were actually found, but even so, flows probably make up less than 1 per cent of the extrusive rocks.

AGE

Leaves and stem fragments are found sparingly in some of the tuffs and volcanic sandstones, and irregular pieces of silicified wood occur rarely in the breccias. Dr. Erling Dorf (written communication) has identified the following species of leaves from a tuff bed at the base of the volcanic series on the west branch of Upper Deer Creek:

Pteris nitida Hollick. One good specimen similar in observable details to specimens of this species from the upper? member of the Chignik formation of Alaska (Montanan age, Upper Cretaceous).

Sequoia reichenbachi (Geinitz) Heer. One excellent specimen indistinguishable from the specimens of this species from the Upper Cretaceous Judith River formation.

Dryophyllum subfalcatum Lesquereux. Two good specimens. Widespread in post-Coloradan Upper Cretaceous.

Rulac quercifolium Hollick. Five good specimens referable to this species from the middle and upper members of the Chignik formation of Alaska (Montanan age, Upper Cretaceous).

Viburnum montanum Knowlton. Five excellent specimens. Widespread in post-Coloradan Upper Cretaceous.

Vitis stantoni (Knowlton) Brown. Three excellent specimens similar in all respects to the type specimen of this species (called *Castalia stantoni* Knowlton) from the Judith River formation. These are quite different from reported occurrences of this species in the higher Upper Cretaceous.

Carpolithus cf. *C. carrollensis* Berry. One excellent specimen very close to this Upper Cretaceous species from the Ripley formation.

? Marchantites, new species.

Dorf (written communication) says: "This combination of species is quite clearly an Upper Cretaceous assemblage of Montanan age. I believe it to be approximately equivalent in age to the Judith River formation of central Montana."

ORIGIN OF EXTRUSIVE SERIES

GENERAL STATEMENT

The Livingston igneous series originated in the intermixing of ejectamenta from a large number of small contemporary vents. During and after the eruptions, the pyroclastic material was modified first by mud flows and later by streams.

VOLCANIC VENTS

Several small vents 1000 to 3000 feet in diameter, filled with agglomerate or breccia occur in the area at the headwaters of the Deer creeks (Fig. 2). The vent agglomerate or breccia in half of these pipes is not much coarser than that in the surrounding areas. However, at least three of the vents were distinguished by fragments averaging 2 to 3 feet in diameter. In one vent blocks more than 6 feet across are numerous. All these vents show vertical jointing and are cut irregularly by intrusive pipes, indicating successive periods of intrusion. It is not possible to tell which part of the extrusive series came from a particular vent, except at the Enos Mountain volcano.

The Enos Mountain volcano (Rouse, *et al.*, 1937, p. 724), a compound dacite breccia plug, is the largest known vent. However, the pyroclastics originating from this vent are limited to 3 or 4 square miles of outcrop area and are only a few hundred feet thick.

EXPLOSION BRECCIAS

The volcanic breccias in the area of the vents, being poorly bedded and having angular fragments embedded in a crystal tuff matrix, appear to be explosion breccias of vulcanian type eruptions. However, the absence of good cone structures indicates post-eruption rearrangement of the ejecta-

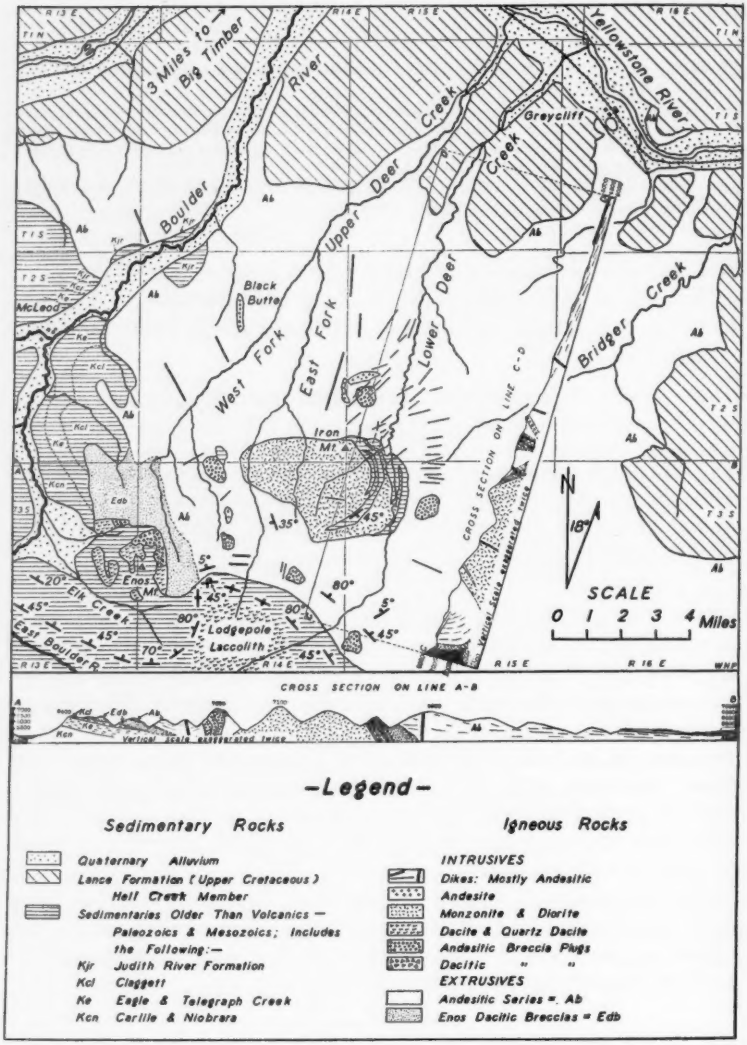


FIGURE 2.—*Geologic map and cross sections of the Deer Creek-McLeod area, Montana*

menta; dry avalanches or mud flows obscured part of the original bedding before consolidation.

MUD FLOWS AND STREAM DEPOSITS

The author believes that the breccias just north of the vents and along Bridger Creek and the Boulder River were deposited chiefly as mud flows because the fragments are heterogeneous and subangular. Vhay (1939, p. 436-437) has suggested that the breccias and tuff-breccias near Nye and just east of the Deer Creek area were deposited chiefly as hot and cold mud flows.

Still farther from the vents to the north and west where volcanic conglomerates and sandstone become numerous, the lenticular beds resemble stream deposits. Deposition by numerous, shifting, overloaded streams on an alluvial slope is suggested. However, even here the matrix of the volcanic sandstones is composed of angular crystal fragments.

INTRUSIVE ROCKS

GENERAL CONSIDERATIONS

The intrusive rocks of the Deer Creek area consist of a small laccolith, many dikes, and a few pipes in addition to the breccia plugs and vents. The Lodgepole laccolith (Rouse, *et al.*, 1937, p. 721-724) belongs to this group (Fig. 2).

IRON MOUNTAIN LACCOLITH

General statement.—An intrusive body 3 to 4 miles across, Gold Hill of Rouse, *et al.* (1937, p. 726), includes Iron Mountain (Fig. 2). This body has raised Paleozoic and Mesozoic sediments, Madison limestone to Cloverly conglomerate inclusive, around its north and east sides to dips of 45° (Fig. 2). A few marginal sills of varying thickness give this body the general form of a Christmas-tree laccolith similar to the Lodgepole laccolith (Rouse, *et al.*, 1937, p. 721-724).

The laccolith is a fine-grained granitic rock varying from augite diorite to hornblende monzonite. The former consists essentially of andesine, augite, magnetite and biotite with accessory apatite, while the latter contains green hornblende, albite-oligoclase, and orthoclase with minor biotite, quartz, magnetite, apatite, and sphene. The biotite is slightly altered to chlorite, and the orthoclase to sericite or kaolin.

Contact metamorphism.—A few small unimportant lenses and masses of magnetite and skarn are located at or near the contact of the monzonite and Madison limestone. Such lenses parallel the bedding of the limestone and reach a maximum length of 20 feet and width of 2 feet. This local contact metamorphism shows chiefly fine-grained magnetite (some lode-

stone), yellowish-brown massive garnet, and fine-grained epidote with minor specularite, wollastonite, and quartz.

Hydrothermal veins.—A number of narrow hydrothermal fissure veins and joint fillings near the center of the intrusive mass have excited prospectors for half a century. These veins, never over 2 to 3 feet wide and a few hundred feet long, consist essentially of calcite and a fine-grained mixture of sericite and quartz with a little pyrite. The latter is reputed to carry traces of gold.

Near these veins the monzonite and crosscutting porphyry dikes show more alteration to chlorite, sericite, and kaolin than elsewhere in the laccolith. This hydrothermal alteration suggests the location of the pipe which fed the laccolith.

DIKES AND PIPES

About 100 narrow vertical dikes radiate from the laccolith. The majority of these are 4 to 10 feet wide, often traceable for many hundreds of feet, and are localized to the north and east of the laccolith (Fig. 2). Northwest of this center, the few dikes present are much wider; the one holding up Black Butte is well over 100 feet wide. A few dikes actually cut the laccolith.

One thick, faulted, semicircular dike north of Iron Mountain dips inward about 45° toward a breccia-filled plug after the manner of a cone sheet.

These dikes are all andesite porphyries with phenocrysts of plagioclase (labradorite and andesine), hornblende, augite, magnetite, and minute apatite crystals in an aphanitic groundmass. About one third are hornblende andesites, one third augite andesites; in the other third hornblende and augite occur in equal amounts. A little biotite is present in about one third of the dikes.

A few small irregular andesitic pipes occur in the breccia plugs as part of the vent structures. An isolated vertical pipe about 1000 feet in diameter is a latite containing about equal amounts of orthoclase and plagioclase with very minor biotite, augite, and magnetite.

STRUCTURE

RELATION TO OLDER ROCKS

The Livingston igneous series overlies with slight angular unconformity the Eagle, Claggett, and Judith River formations as exposed along the Main Boulder and East Boulder River valleys. This unconformity represents a surface of low relief and dips gently northward from the vicinity of Enos Mountain reaching stream level a few miles north of McLeod in the Boulder valley.

On the east side, in the Stillwater area (Fig. 1), the same angular unconformity was mapped by Wilson (1936, p. 1170), Vhay (1939, p. 434), and Stow (written communication).

Several intrusives (Vhay, 1939, p. 434) outcrop south of the volcanic series, principally the Lodgepole laccolith (Foote *in* Rouse, *et al.*, 1937, p. 721-723). These bodies have tilted Paleozoic and Mesozoic sediments to near vertical. Later the main Beartooth thrusting has further deformed these beds. Some of the pyroclastics have also been tilted particularly near the headwaters of Lower Deer Creek. Here the volcanics are nearly conformable with Upper Cretaceous beds of Coloradan or Montanan age. However, some of the extrusive beds overlie unconformably the upturned sediments particularly near the headwaters of Upper Deer Creek where nearly horizontal breccias overlie vertical and overturned Cretaceous beds, Cloverly and younger. Thus some breccias postdate the intrusion of the Lodgepole laccolith, although others are older and were tilted by the doming of this laccolith. The Lodgepole laccolith was intruded prior to Beartooth thrusting (Rouse, *et al.*, 1937, p. 722).

RELATION OF YOUNGER ROCKS

Northward and northeastward the gently north dipping volcanic series is overlain conformably by Lance sediments. The bed immediately overlying the volcanics from the Boulder River area near Big Timber around to the Stillwater River area has been identified as the Hell Creek member of the Lance by Stow (written communication).

STRUCTURES WITHIN THE VOLCANIC SERIES

The lack of good bedding in the volcanic series of the vent area, due to dry avalanches and mud flows, has made it impossible to identify any actual cone structures around the various vents and difficult to distinguish dips of deposition from deformational structures. Fortunately, sufficient bedding was present to indicate the relationships at the south edge of the Deer Creek area.

The relation of the breccias to the Iron Mountain laccolith is rather obscure. Certainly much of the breccia is later than this laccolith and overlies it and the upturned sediments unconformably. Some steeply dipping but poorly bedded breccias near the southwest side of the laccolith suggest that some pyroclastic material was domed by the emplacement of the Iron Mountain laccolith.

Two angular unconformities have been recognized within the volcanic series. The Enos Mountain dacitic breccias are overlain by andesitic breccias with a slight angular unconformity. At least one similar unconformity exists within the andesitic beds, best seen north of Iron Mountain and east of Lower Deer Creek where truncated breccias dipping 45°

NE. are overlain by nearly horizontal breccias; this steep dip may be partly depositional. Periods of eruption seem to have been separated by periods of considerable erosion and some tilting and deformation.

SUMMARY OF CONCLUSIONS

GENERAL CONSIDERATIONS

Geographically and genetically the Livingston pyroclastic and intrusive rocks are a distinct unit of the western Montana upper Cretaceous volcanic field.

The pyroclastic material was exploded from a large number of small vents and was modified by mud flows and streams.

Intrusive rock types include andesite, latite, dacite, diorite, and monzonite. Augite and hornblende are almost equally abundant, and biotite in small amounts is characteristic. Quartz is absent or rare. The pyroclastics contain fragments of these same rocks and minerals, though andesite greatly predominates.

SUMMARY OF EVENTS

(1) The first extrusive activity in the Deer Creek area occurred in Judith River time as indicated by leaf fossils. The Enos Mountain dacitic vent was one of these early extrusive centers. As eruptions occurred in some areas, Judith River sediments were being deposited simultaneously nearby.

(2) Slight erosion of dacitic pyroclastics and Judith River sediments.

(3) Widespread andesitic breccias were deposited unconformably on dacitic breccias and Judith River sediments. Periods of volcanic activity alternated with periods of erosion and redeposition.

(4) Intrusion of the Lodgepole and Iron Mountain laccoliths domed Paleozoic and Mesozoic sediments as well as the earlier breccias.

(5) Widespread erosion of the earlier vents and pyroclastics uncovered the Iron Mountain intrusive rock. The products of this erosion were deposited to the north and west as tuffaceous sandstones and volcanic conglomerates.

(6) Extrusive activity occurred from a large number of small vents. This later material was deposited unconformably on the tilted and beveled edges of the earlier pyroclastics.

(7) Intrusion of the majority of the andesitic dikes most of which cut the later breccias.

(8) Volcanic activity had ceased by Lance time. The Hell Creek member of the Lance overlies conformably the volcanic conglomerates, tuffaceous sandstones, and breccias to the north and east.

(9) The main Beartooth thrusting followed this volcanic activity but did not affect the volcanic series appreciably.

REFERENCES CITED

Iddings, J. P., and Weed, W. H. (1894) *Livingston, Montana,* U. S. Geol. Survey, Geol. Atlas, Folio no. 1.

Rouse, J. T., *et al.* (1937) *Petrology, structure and relations to tectonics of porphyry intrusions in the Beartooth Mountains, Montana,* Jour. Geol., vol. 45, p. 717-740.

Stone, R. W., and Calvert, W. R. (1910) *Stratigraphic relations of the Livingston formation of Montana,* Econ. Geol., vol. 5, p. 551-557, 652-669, 741-764.

Vhay, J. S. (1939) *Some features of the Livingston formation near Nye, Montana,* Am. Geophys. Union, Tr., p. 433-437.

Weed, W. H. (1893) *The Laramie and the underlying Livingston formation in Montana,* U. S. Geol. Survey, Bull. 105.

Wentworth, C. K., and Williams, Howel (1932) *The classification and terminology of the pyroclastic rocks,* Nat. Res. Council, Bull., no. 89, p. 19-53.

Wilson, C. W., Jr. (1936) *Geology of the Nye-Bowler lineament, Stillwater and Carbon Counties, Montana,* Am. Assoc. Petrol. Geol., Bull., vol. 20, p. 1161-1188.

National Bureau of Standards, Washington, D. C.
Manuscript received by the Secretary of the Society, July 28, 1941.
Presented before the Geological Society, December 30, 1939.
Project Grant 290-39.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 53, PP. 1187-1210, 4 FIGS. AUGUST 1, 1942

TIME RELATIONS IN OCEAN SEDIMENTS

BY C. S. PIGGOT AND WM. D. URRY

CONTENTS

	Page
Abstract	1187
Foreword	1188
Acknowledgments	1188
Introduction	1189
Method	1190
Time intervals in ocean-bottom cores	1194
North Atlantic cores	1194
Globigerina ooze from the Caribbean Sea	1199
Red clay from the Pacific Ocean	1200
Rate of deposition of ocean sediments	1201
General statement	1201
Influence of distortion	1201
North Atlantic sediments	1203
Caribbean Sea sediments	1204
Red clay from the Pacific Ocean	1205
Compaction	1205
Deposition in terms of mass	1207
Summary	1207
Notes on the cores of ocean-bottom sediments	1208
References cited	1208

ILLUSTRATIONS

Figure		Page
1.	Variation of the radium concentration below the surface of the ocean bottom	1191
2.	Chronology of sediments in the North Atlantic	1195
3.	Changes in the abundance of Foraminifera with time in the Caribbean Sea	1200
4.	Variations in the rate of deposition of ocean sediments in the past	1202

ABSTRACT

Nonequilibrium between the radioactive elements in the surface of the ocean bottom provides a method of measuring time in the past 300,000 years. The principles of this method are not to be confused with those underlying the radioactive methods of determining geologic time measured in millions of years. The well-preserved geological and biological history of the ocean basins has been worked out for many of the cores of ocean sediments procured by Piggot, and the above method of measuring time intervals in years has been applied to these cores. The results indicate: (1) that the effects of glaciation on the continents are contemporaneous with equivalent effects on the type of deposit in the ocean bottom; (2) that the effects of glaciation on the type of ocean sediment are widespread, extending in the northern hemisphere at least to the Caribbean Sea; (3) that within short intervals of time there is, in the ocean bottom, a considerable variation in the type of deposit attributa-

ble directly or indirectly to climatic changes which, probably because of the continuous effacement of the evidence on land, has not been reported in studies of continental glaciation.

This method cannot compete in accuracy with the use of varved clays, but it possesses the advantages of (1) application to a considerable area of the earth's surface, (2) reference to the present time, and (3) applicability to at least half of the Pleistocene epoch.

The rate of deposition of ocean sediments is treated as a subsidiary problem. The obvious step from the determination of time intervals in a core to the rate of deposition of the sediment represented by the core is complicated by the present incomplete knowledge of the distortion of the sediment in the process of taking the sample. The results in general agree with previous estimates by other methods, some of which, however, are open to the same criticism. It is possible to discuss within limits the variations of the rate of deposition of ocean sediments in the past.

FOREWORD

The lithology, biology, and chemistry of cores of ocean-bottom sediments have been described in the literature. The comprehensive examination by Bradley and associates (Bradley *et al.*, 1940) of the geology, chemistry, and biology of a suite of North Atlantic deep-water cores obtained by Piggot in 1936 constitutes a great advance in such researches, primarily because the cores were considerably longer than those previously available, and because of the wealth of detail that has been accumulated by a number of specialists working on identical material.

The concentrations of the radioelements in these cores have been determined, and the radioactive relations thus established offer a method of dating any horizon in these vertical sections of the ocean bottom. The geological and biological studies indicate several important datum planes, which provide added interest in an objective method by which core lengths may be converted into time intervals.

Measurements of the time intervals have been made in some cores not included in the North Atlantic group. The Foraminifera from one of these cores from the Bartlett Deep in the Cayman Trough have been studied by Cushman (1941), and his findings stimulated the present work on this core.

In order that a given date may have significance, some event must exist; the reports, to which reference is made above, provide many such events.

ACKNOWLEDGMENTS

Thanks are extended to Dr. Joseph A. Cushman for helpful discussion and communication of unpublished results of foraminiferal studies in these cores; to Dr. Roger Revelle of the Scripps Institution of Oceanography, La Jolla, California, for supplying many specimens of a core of Red clay; and to Dr. Juul Hvorslev for helpful criticism and discussion of the problem of distortion in the process of taking samples of the ocean bottom.

INTRODUCTION

Heretofore there has been no direct method of dating the many events that have left their records undisturbed in the ocean bottom.

Rate of deposition of ocean sediments has been estimated by assuming a date for an observed event, as, for example, the estimate that the accumulation of postglacial warm-water ooze began in the equatorial Atlantic 20,000 years ago. By this method Schott (1935, 1939) calculated the average rates of accumulation of the three main types of ocean sediments. Lohmann (1909) based similar estimates on the population and rate of growth of the Coccolithophoridae, and Twenhofel (1929) estimated the amount of inorganic material deposited on the sea floor remote from the coasts. The estimates by Schott and the measurements reported here indicate that the fluctuations over any considerable area are so great as to invalidate the timing of events at one place in the ocean by the data obtained from another locality.

It is now known that the core length does not bear a one to one, or uniform, ratio to the depth of the sediment sampled (Piggot, 1941; Emery and Dietz, 1939; 1941), and determinations of rates of deposition of the sediments are accompanied by uncertainties that are difficult to evaluate. This applies to such estimates as have been made previously and also to the values given here. These uncertainties do not exist in the present method of dating a particular event, the evidence for which is found at a given core depth, and to the authors this seems to be the more immediate problem.

Any direct observations of the rate of deposition of sediments as made by Murray and Peake (1904) are subject to considerable experimental error and pertain only to the present time and place of observation. Lack of knowledge of the rate of deposition of all classes of sediments during the past seriously hinders the use of the geological sediment column as a time scale (Schuchert, 1931). For this reason, the physical properties of the radioelements, whose rates of disintegration are invariant, have been used to estimate geologic time. Since helium and lead are the ultimate stable products of the radioactive disintegrations of uranium and thorium, the former elements have become associated with two established methods of determining geologic time (Lane, 1924-1941). While the principles of these methods are simple, the applications are subject to inherent uncertainties.

For measuring recent intervals of time in the Pleistocene, some properties of the radioelements can be employed which do not depend upon the accumulation of the stable end products. This results in a method that is largely free from the uncertainties of the "helium" and "lead" techniques, which in any case are not applicable to the immediate past.

The method has been aptly termed by Lane (1937) the "Per cent of Equilibrium Method," but its principle has been recognized since the early days of the science of radioactivity. It has received scant attention because it is limited to the past half million years.

METHOD

Radium is the third element of importance in the series of disintegrations by which uranium is transformed into lead. If a geological specimen has remained chemically undisturbed throughout the Pleistocene or longer, the elements in this series will be in "radioactive equilibrium," and the relative amounts of each will be fixed and independent of time. If, in the course of formation of a more recent specimen, radium should be segregated from its progenitors, its concentration in such a specimen will decrease exponentially with a half-life of 1700 years. This simple example of the method will rarely apply in nature, although Schlundt (1934) apparently found it in the tufas of Yellowstone Park.

A more complicated case exists if the uranium is completely segregated from its daughter and granddaughter elements ionium and radium. For practical purposes, the uranium may be considered as an inexhaustible reservoir. It cannot produce radium directly; it must first disintegrate into ionium, which has a half-life of 82,000 years. Consequently, the radium content will increase with time but at a considerably slower rate than it decreases in the simple case treated above. This provides a more promising means of measuring time in the Pleistocene than that depending upon the disintegration of radium by itself, because after 300,000 years there is still a detectable difference between the radium concentration and the ultimate constant value that it will attain when radioactive equilibrium has been established. The theory of this case has been treated by Kovarik (1931). Such clean-cut separations seldom occur in nature, and their existence would be difficult to substantiate.

Cores of ocean-bottom sediments, which rarely are disturbed by tectonic or other physical forces, provide a means of following the concentrations of the radioelements as these change with time. Such studies (Piggot and Urry, 1939; 1941; 1942; Urry and Piggot, 1942) have revealed that all three of the radioelements mentioned above are being incorporated in the structure of the ocean sediments at any certain place. The ratios of the concentrations of uranium, ionium, and radium being deposited differ greatly from the fixed ratios in radioactive equilibrium, and thus there is a complicated variation of the radium content with depth below the surface of the sediment. The ocean sediments as laid down contain very little uranium, considerably more ionium

than is equivalent to this uranium, and, in most cases, less radium than that appropriate to equilibrium with the ionium.

In Figure 1, curve A illustrates the change of the radium content with time. The ionium first produces radium at a rate in excess of that at which the radium decays, therefore the radium content increases until these rates are equal. The time required to accomplish this is estab-

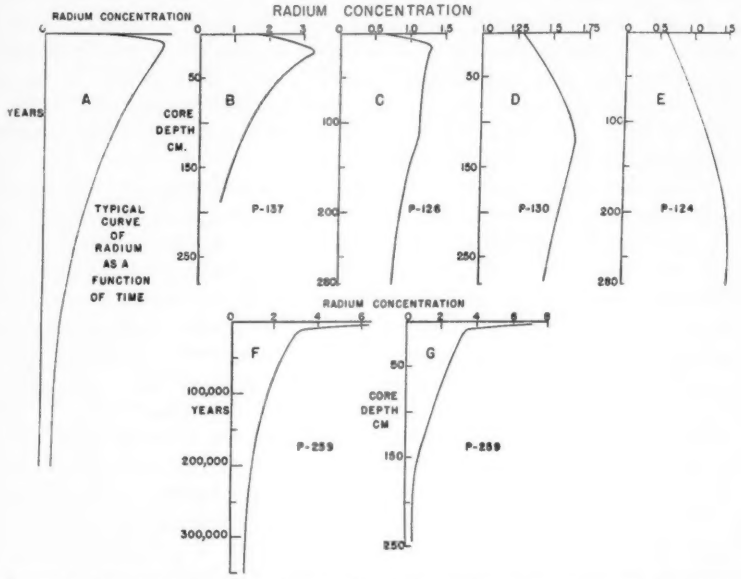


FIGURE 1.—*Variation of the radium concentration below the surface of the ocean bottom*

Curves A and F express the radium concentration as a function of time, calculated from the equations of radioactivity for two sets of initial conditions. Curve F is calculated for the actual initial conditions found experimentally (curve G) in core P-259. Curve A is a typical curve which must be calculated for each of the specific cases represented by the experimental curves B, C, D, and E. The radium concentration is expressed in 10^{-12} g. radium per cc. of the original sediment—*i.e.*, it is the radium content in a sample of standard cross section (1 sq. cm.) and unit depth (1 cm.).

lished by the laws of radioactive disintegration, and it is dependent upon the ratios of the initial amounts of radium, ionium, and uranium. The whole of curve A is equally dependent upon these ratios, and it must therefore be calculated for each core. After the attainment of equilibrium with its parent ionium, the radium concentration is controlled by the decay of the former element. The ionium concentration decreases because of insufficient uranium to support it. Finally, the number of atoms of ionium disintegrating will become just equal to the number

of new ionium atoms produced from the small amount of uranium present, and radioactive equilibrium among all three elements will be established. This is illustrated in Figure 1, curve A, which shows the attainment of an essentially constant radium content after the lapse of 300,000 to 400,000 years. Curve F in Figure 1 is the theoretical curve of radium as a function of time for a core of Red clay, and curve G is the experimental curve of radium content as a function of depth in the same core. This Red clay differs from the other sediments examined in that radium was deposited in excess of that which is appropriate to equilibrium with the ionium deposited. Why this should be so is not yet known, but chemically and biologically the Red clay differs greatly from the calcareous blue muds and Globigerina oozes which comprise the other examples in Figure 1.

The analysis of each curve in order to convert units of core depth into time intervals necessitates superposing the appropriate curve A on the experimental curve. The mathematical treatment of such complex mixtures of the radioelements and the method of accomplishing this conversion have been given by Urry (1942). It is not possible to superpose the curves graphically, and the somewhat lengthy mathematical process cannot be discussed here. However, the procedure may be studied qualitatively from Figure 1. Curve B needs little explanation because its profile is so similar to that of the theoretical curve A. In curve C, the more gradual decrease in the radium content indicates that a much shorter period of time was required to deposit 280 cm. of core P-126 in the North Atlantic than to lay down the 190 cm. of core P-137 in the Cayman Trough of the Caribbean Sea. Still shorter periods of time are involved in cores P-130 and P-124 where the maximum radium content is about half way down, and near the bottom of each core, respectively. Curves B, C, D, and E therefore represent decreasing intervals of time required for the deposition of the sediments represented by the respective cores. Changes in the rate of deposition are shown by a departure of the experimental curve from the appropriate theoretical curve. Failure of the radium to decrease normally for a certain portion of the core indicates an accelerated rate of deposition for a certain time, whereas an abnormal decrease in the radium content indicates a retarded rate of sedimentation. For instance, curve C, for P-126, indicates an abnormality in the rate of deposition between 80 and 120 cm. from the top of the core, which will be discussed in the next section.

The experimental curve G for the Red-clay core, P-259, demonstrates the attainment of complete radioactive equilibrium as shown by the constancy of the radium content below about 160 cm. In this region the radium content provides a measure of the uranium content which

for all practical purposes is constant over the interval of time represented by any of the cores. The initial ionium content can be determined from the slope of that portion of the curve which represents the decay of the ionium. Thus it is possible to acquire a complete knowledge of the relative amounts of uranium, ionium, and radium at any time merely by measuring the relative concentrations of radium at a series of depths in a core. It is not even necessary to refer the determinations to a radium standard. This has been done, however, because the radium determinations are of use for other purposes besides the dating of events in the sediments.

The cores represented by curves B, C, D, and E, although as long as P-259, are far from representing the interval of time required to establish radioactive equilibrium. In such cases the uranium must be measured by an independent method. Fortunately, a knowledge of the uranium content is imperative only when the core represents a considerable fraction of the time required to attain equilibrium, as is the case with P-137. An independent measurement of the uranium content of this core has been made by Urry (1941). The uranium content is quite unimportant in the analyses of curves D and E, in Figure 1.

It must be emphasized that the conversion of the core depth in centimeters, into time in the past—*i.e.*, the superposing of the theoretical curve A on the experimental curve B—requires the assumption that the sediment has acquired a constant concentration of the radioelements from the sea water or other sources throughout the period of its deposition. This need only be true of the deposit at the location of the core. Quite appreciable short-period fluctuations will cause but little error provided these smooth out within 100 years or more of sedimentation. The following facts support this assumption. In cores of uniform composition the experimental curves are quite smooth. The uranium content of core P-137 is constant throughout the length of the core, within the limits of error of the measurements (Urry, 1941). If the variations of the radium content, as shown in the experimental curves, are due to unknown causes quite unrelated to the established laws of radioactivity which give curves A and F, then the unknown causes must produce these variations according to equations analogous to those governing radioactive decay. The authors have examined this possibility but find no basis for such a contention.

Core P-126 contains several zones of glacial marine deposits where the character of the sediment is different from the warm-water zones (Bradley *et al.*, 1940). As might be expected, the radium concentrations in these coarser sediments depart, although not seriously, from the curve drawn through the points for the normal sediment of the core. For

this reason, the analysis of the cores from the North Atlantic must be more detailed than those of the sediments from lower latitudes (P-137, P-259), which are known to be more uniform in character.

TIME INTERVALS IN OCEAN-BOTTOM CORES

NORTH ATLANTIC CORES

The complete suite of ten cores obtained by Piggot on board the cable ship LORD KELVIN from the North Atlantic has been examined lithologically and geologically by Bramlette and Bradley (Bradley *et al.*, 1940, Pt. 1), and the Foraminifera have been studied in detail by Cushman and Henbest (Bradley *et al.*, 1940, Pt. 2). The three cores from this suite chosen for the preliminary radioactive investigations showed the greatest difference in the relative depths of the zones as established by the lithological and foraminiferal studies. The disposition of the zones has been plotted in Figure 2 against the core depth. The key to the shading is that employed by Bramlette and Bradley (Bradley *et al.*, 1940, Pl. 3), with the exception of the zones of anomalous deposits. The numbers opposite the lines defining the zones or the temperature dots give the intervals of time in years measured from the top of the core as zero.

While it is unnecessary to redescribe the features of each zone, which can now be assigned a date instead of a depth in the core, attention may be drawn to a few important points.

That the time interval represented by the North Atlantic cores varied considerably was obvious even before the results described here were available. This was demonstrated not so much by the presence and number of glacial zones [which need not be contemporaneous, if Bramlette and Bradley's (1940, p. 7-9, Fig. 2) interpretation of their formation be accepted] as by the position of the well-defined base of the zone of alkalic volcanic glass shards which is part of the uppermost warm-water zone in all the cores except No. 11. Further evidence was provided by a lower volcanic zone in all the cores [except P-124(3)] west of the Mid-Atlantic Ridge. This zone was not penetrated even at greater core depths east of the Ridge. The second volcanic zone is shown in core P-126(5) in Figure 2. In core P-124(3) no glacial zone was penetrated, and the base of the upper zone of volcanic shards was not reached.

These shards are assumed to have originated in explosive volcanic eruptions and to have been transported initially through the air. They are more numerous at the bottom of the layer, which is sharply defined so that their deposition implies contemporaneity over large geographical areas. This is confirmed by the occurrence of the base of the upper ash

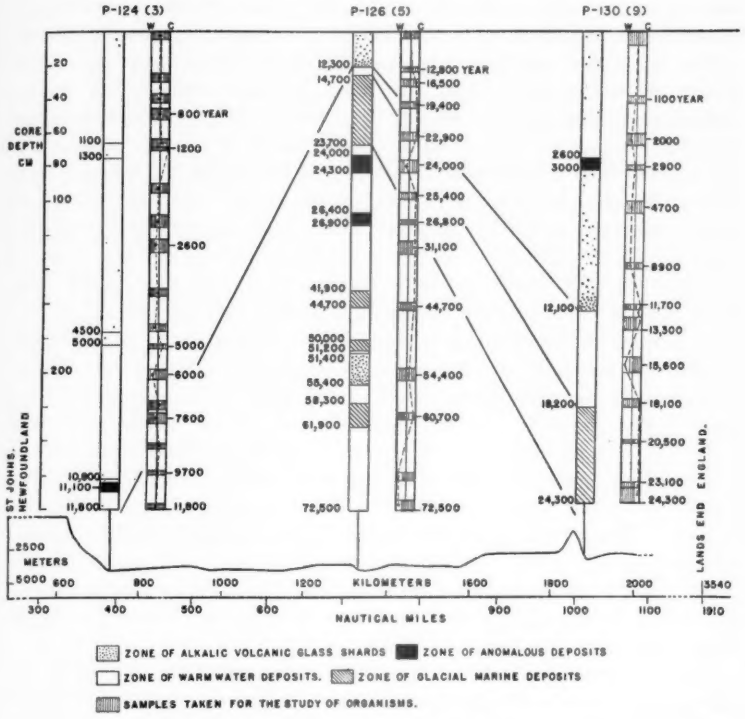


FIGURE 2.—*Chronology of sediments in the North Atlantic*

The dates of various events (present time = 0) which have left evidence of their occurrence in the ocean sediments of the North Atlantic. The vertical scale is in centimeters of core depth; the horizontal scale in distances. Each core is represented by two profiles. The left-hand profile shows the lithological events (Bradley *et al.*, 1940, Pl. 3), and the right-hand profile shows the past fluctuations in the temperature of the surface water at the site of each core (Bradley *et al.*, 1940, Pl. 4). The dot in each shaded block indicates qualitatively the deviations from the present temperature. The station numbers in parenthesis after each laboratory core number are the same as those used by Bradley (1940).

The numbers against the lines defining the zones or opposite the temperature dots give the intervals of time in years measured from the top of the core as zero. The quotation of time intervals to the nearest 100 years must not be taken too literally. Such accuracy of statement is probably warranted in the case of P-124 (3) and for P-130 (9) above the zone of glacial marine deposits. Some of the figures for core P-126 (5) are within this degree of accuracy, but little significance should be attached to differences of less than 1000 years below the second glacial zone in this core.

zone at 12,000 years in both P-126(5) and P-130(9). The fact that core P-124(3) represents no more than 12,000 years places it entirely in the upper volcanic zone, as is indicated by the lithology. According to the dating of the second volcanic zone in core P-126(5), similar ash

was deposited between 51,000 and 55,000 years ago and was followed by conditions which gave rise to 1000 years of glacial marine deposition.

The zones that are dated in core P-124 are three thin and rather widely spaced layers of less limy mud (Bradley *et al.*, 1940, p. 5). Cushman and Henbest (Bradley *et al.*, 1940, Pl. 4) find some indications of colder surface waters during the deposition of these clay zones. The anomalous zone of terrigenous silt at 11,000 years presumably was caused by slumping (Bradley *et al.*, 1940, p. 16). Between 2000 and 4500 years ago, and again at about 6000 years ago, in core P-124(3), there is some indication that the surface water was warmer than at present. Core P-130(9) indicates that a warmer period existed from 2000 to 9000 years ago, and also about 16,000 years ago.

Four zones of glacial marine deposits were penetrated in core P-126(5). This is the maximum number found in any of the cores. All four zones occur within the past 62,000 years. These zones in core P-126(5) have been plotted against time in Figure 3, and all are contemporaneous with the Würm glacial age. The chronology of the Pleistocene for the Alps given in Figure 3 is that of Penck and Brückner (1901-1909). Kay (1931) has made a detailed study of the chronology of the Pleistocene in Iowa. His results are reproduced in Table 1 because of their bearing on the problem in the North Atlantic and elsewhere. The glacial ages in Iowa are brief because they represent only the most southerly extremities of the ice sheets, and therefore the time when the respective glaciations were at a maximum.

Kay's values are plotted at the top of Figure 3. The earliest glacial marine zone in core P-126(5) was deposited between 58,000 and 62,000 years ago. These dates are in reasonable agreement with Kay's proposal of a separate Iowan age between 55,000 and 58,000 years ago. Bramlette and Bradley (Bradley *et al.*, 1940, p. 11) favor the interpretation that all four glacial marine zones are substages of the Wisconsin stage. They do not mention an Iowan glaciation, either as a separate stage, or as a substage of the Wisconsin, but they discuss the early and late substages of the Wisconsin as proposed by Chamberlin and Salisbury (1906). The lower zone of volcanic ash above the earliest glacial marine zone may offer some justification for setting off the Iowan glaciation from the Wisconsin substages. Moreover, it is only in this interglacial marine zone in core P-126(5) that the temperature conditions approach those of the present day. This is substantiated by the temperature charts for the other cores 4, 6, and 7 (Bradley *et al.*, 1940, Pl. 4).

Contemporaneity of the earliest glacial marine zone with Kay's Iowan age implies contemporaneity of the Peorian loess with the lower volcanic ash zone. Kay (1931, p. 443) writes, "This loess was apparently laid

down shortly after the deposition of the Iowan drift and not long before the deposition of the early Wisconsin drift." The base of the ash zone coincides with Kay's close of the Iowan age at 55,000 years.

It is interesting to note a considerable length of core between the last and next to last glacial marine zone indicated as a warm-water deposit

TABLE 1.—*Minimum duration of the Pleistocene period in Iowa*

(After G. F. Kay)

Interglacial age	Years	Glacial age
	0	
(Post late Wisconsin)		
	25,000	Late Wisconsin
	28,000	
Peorian		
	55,000	Iowan
	58,000	
Sangamon		
	178,000	Illinoian
	187,000	
Yarmouth		
	487,000	Kansan
	495,000	
Aftonian		
	695,000	Nebraskan
	705,000	

in core P-126(5), particularly since estimates of the time required for the recession of the continental ice fronts such as that of Antevs (1928) would indicate that this section of the core should be contemporaneous with the peak of the last glaciation. The temperature chart of Cushman and Henbest (Bradley *et al.*, 1940, Pl. 4), reproduced here in Figure 2 for the cores under consideration, indicates a continuation of cold-water conditions in this interval, and Bramlette and Bradley (Bradley *et al.*, 1940, p. 8-9) find the texture and composition of this zone to be intermediate between glacial marine deposits and foraminiferal marl, from which they infer a repeated migration of the limit of the drift ice back and forth across the site of these cores. The time represented by this zone in core P-126(5), when judged by its depth, is misleading. This is true of many core samples because of the distortion of the sediment introduced by the sampling procedure, as will be discussed under "Rates of Deposition." In this case much of the apparently long duration of the intermediate zone is due to two zones of anomalous deposits represented by the Bradley samples Nos. B-53 and B-55 (Bradley *et al.*, 1940,

Pl. 3). The upper zone, B-53, is 10 centimeters deep, and the lower zone, B-55, is about 8 centimeters deep. The radium content of both these zones is about the same and is considerably less than that indicated by the curve C in Figure 1 for the respective positions in the core; in fact, the radium content is comparable with that of the material between the earliest zone of glacial marine deposit and the lower volcanic ash zone. The time intervals as measured for these anomalous zones are extremely small and, within the limits of error, might be zero. This accounts for the curious shape of curve C in Figure 1. The accuracy of the method is insufficient to justify the conclusion that these anomalous zones originate by slumping, and the inference that the source of the slumped material was the sediment lying now at about 220 cm. in the core is unwarranted. Nevertheless, there is strong evidence for such a mechanism for the origin of the upper anomalous zone and very suggestive evidence for a similar origin of the second anomalous zone (Bradley *et al.*, 1940, p. 15-16). Bramlette and Bradley also suggest that the sediments of the lower zones of this core are a possible source of the material on the basis of lithological and temperature similarities.

Core P-130(9) provides an opportunity for detailed study of the features presented in the top 70 centimeters of core P-126(5). The contemporaneity of the boundary marking the presence of volcanic shards has been considered already. It is disconcerting to find that the deposition of the last glacial marine deposit ceased 3500 years earlier in this region than in the neighborhood of core P-126(5), although Cushman and Henbest find evidence for a brief cold-water period between 12,000 and 13,000 years ago, as shown in Figure 2. It is possible that the dates given for the lower portion of core P-130(9) are in error because the radium values for material in the glacial zone had to be given some weight. As previously mentioned, such values are subject to the influences of the source of the detrital material. In core P-126(5) there is evidence of four distinct periods of glacial marine deposition in 47,000 years, which requires considerable movement of the southern limit of the drift ice within a single glacial age, or two concurrent ages if Kay's Iowan be accepted. In the zone of intermediate character in core P-126(5), between 24,000 and 42,000 years, these northerly and southerly movements must have occurred in fairly rapid succession. Such a picture has been illustrated for a meridional section by Bramlette and Bradley (Bradley *et al.*, 1940, Fig. 2). The southerly limit of the drift ice, however, possibly varied at different longitudes at different times, so that the glacial marine deposits in the North Atlantic may not be completely contemporaneous.

GLOBIGERINA OOZE FROM THE CARIBBEAN SEA

Three cores from the bottom of the Cayman Trough in the vicinity of the Bartlett Deep were obtained by Piggot on board the ATLANTIS in 1937. The sediment is Globigerina ooze. The Foraminifera have been studied by Cushman (1941). Many interesting inferences have been drawn from these studies which make it more desirable to plot Cushman's relative abundance curves for the Foraminifera, as a function of time instead of core depth. It is first important to ascertain the correlation of these curves, which are believed to indicate swings of warmer and colder surface temperatures. The curves for the three cores P-135, P-136, and P-137 (Cushman, 1941, Figs. 5-10) show no correlation when plotted as a function of core depth. The method of dating cores described here offers an explanation. Each core was deposited in a very different interval of time. The radioactive measurements (Piggot and Urry, 1942) have not been analyzed completely, and the final results will be reported later. The results for P-137 are of interest. Cushman (1941) draws attention to the fact that ". . . the swings shown in core P-136 (Text Figs. 6, 7, 8), in particular, may be coincident with known glacial and interglacial periods." Cushman chose P-136 as probably representing the longest time interval on the basis of Schott's mean rates of sedimentation in the ocean and because of the well-marked periodicity of the abundance curves. The radioactive measurements indicate that the sediment represented by core P-137 required 300,000 years for its deposition, which is more than twice the time required for the deposition of the sediment represented by core P-136, although it is of similar length.

In Figure 3, some of Cushman's abundance curves are plotted as a function of time. The relative abundance of *Globorotalia menardii* has been studied by Schott (1933) as an index of surface-water temperatures comparable to those now prevailing in the equatorial Atlantic and therefore has been included. The duration of the Pleistocene after Penck and Brückner (1901-1909), and the duration of the Pleistocene ages in Iowa after Kay as given in Table 1, are plotted on the same time scale. The North Atlantic core P-126(5) is plotted on the same scale for the purpose already mentioned. There appears to be some correlation between the cold- and warm-water swings, as exhibited by the foraminiferal remains in core P-137, and the Pleistocene ages of glaciation. A more detailed examination of the Foraminifera in this core, between the positions corresponding to 55,000 and 100,000 years, was undertaken by Doctor Cushman at the writers' request, after the publication of his earlier results and after the time scale had been assigned. These additional data were required mainly to establish the correlations between cores P-136 and P-137, but it is interesting to note that cold-water zones,

which were not evident from the first examination, but which were predictable from the present results for the North Atlantic cores and from Kay's results, have been found at about 55,000 and 65,000 years ago. However, the additional data indicate the existence of a brief period of colder surface water about 80,000 years ago, which has no apparent counterpart in the chronology of continental glaciation.

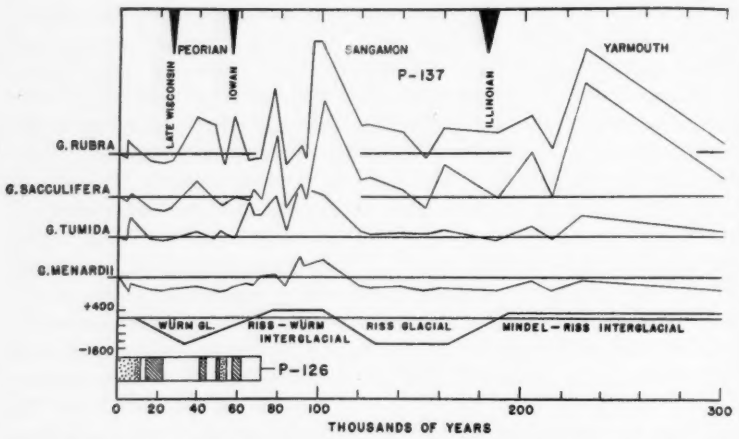


FIGURE 3.—*Changes in the abundance of Foraminifera with time in the Caribbean Sea*

The relative abundance of four of the pelagic species of Foraminifera studied by Cushman in core P-137 plotted against time. On the same abscissa of time are plotted the chronology of the Alpine glaciations after Penck and Brückner, in terms of snow-line level in meters, the chronology of the Iowan glaciations after Kay (at the top of the diagram), and the chronology of the various zones in core P-126 (5) with the same key of shading as in Figure 2.

RED CLAY FROM THE PACIFIC OCEAN

Dr. Roger Revelle supplied specimens from a core of brownish red clay. The radium measurements on this material (Geophysical Laboratory No. P-259) were useful in illustrating the method earlier in this paper. The core was 246 cm. long, the upper 160 cm. of which represent 300,000 years. Below this, the method of dating becomes inaccurate because, as already explained, the radium concentration has practically reached a constant value in radioactive equilibrium with the uranium and ionium. If the rate of sedimentation at the depth of 140-160 cm. be applied to the sediment in the bottom of the core, the whole core represents a thickness of sediment requiring from 650,000 to 800,000 years for deposition. The wide limits are necessary because the distortion of the core during the process of sampling is difficult to appraise numerically.

RATE OF DEPOSITION OF OCEAN SEDIMENTS

GENERAL STATEMENT

In the previous section the authors have distinguished between the core and the sediment in the ocean bottom that the core represents. The length of the core is not the depth of the sediment that has been penetrated (core penetration), nor is this core penetration always equal to the instrument penetration—*i. e.*, the depth of penetration of the coring instrument.

From test sampling conducted by Piggot (1941) in the varved-clay pits near Hartford, Connecticut, it has been possible to establish the relations of these various depths. The data obtained in sampling the varved clays must be applied with caution to the cores of ocean sediments. A comparison of the mechanical properties of the varved clays with those of the cores discussed here indicates that the distortion of the ocean-bottom cores is perhaps somewhat less than that of the varved clays. The calculation of the rates of deposition of ocean sediments from the time intervals in ocean-bottom cores requires a knowledge of the depth of sediment represented by a given length of core. For the upper third of a core of varved clay this ratio is close to unity, so that the uncertainties involved in applying the data for the varved clays to ocean sediments probably do not exceed the errors of the measurements of the time intervals. However, this ratio of undisturbed sediment to core progressively increases and at the bottom of the cores of varved clay attains a value of 4 to 5.

INFLUENCE OF DISTORTION

The rates of deposition are presented here for the limiting cases. In Figure 4 is shown the rate of deposition of ocean-bottom sediments in the North Atlantic, as revealed by three cores from locations several hundred miles apart, the rate of deposition of a Globigerina ooze in the Caribbean Sea, and of a Red clay in the Pacific Ocean. The solid lines give the rates if there is no distortion of the sediment in obtaining the core, and the dashed lines give the rates when the depth of sediment is calculated from the core length with the distortion factors found for the varved clays. The true rate of deposition cannot be less than that indicated by the solid lines and is probably nearer to, but perhaps somewhat less than, that given by the dashed lines. This uncertainty, caused by a lack of knowledge of the distortion during sampling, applies also to the cores used by Schott (1935) to determine his oft-quoted estimates of the mean rates of deposition of the various types of ocean sediments. Most of Schott's cores were taken with a 20 mm. liner, which would introduce greater distortions than would the 50 mm. liner used by the authors.

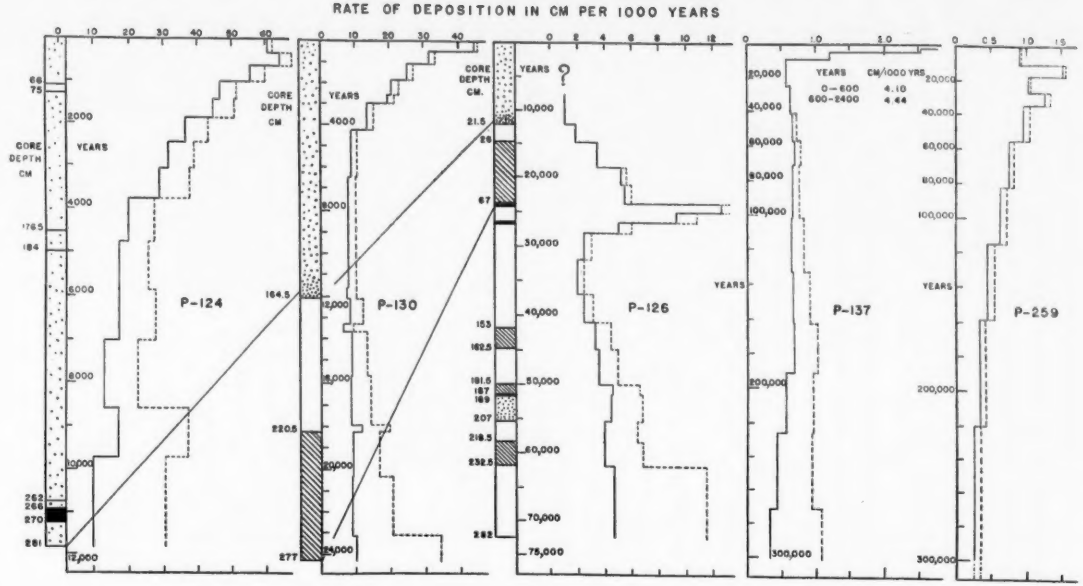


FIGURE 4.—*Variations in the rate of deposition of ocean sediments in the past*

Some rates of deposition of ocean sediments in centimeters per 1000 years plotted against time. Solid lines are from values based on an assumption of no distortion of the sediment when sampling, dashed lines are from values based on the assumption that the distortion factors are the same as in the varved clays of Connecticut. The key to the profiles of the North Atlantic cores is the same as in Figure 2. Numbers to the left of each profile are core depths for important levels. The cross-reference lines for P-124, P-130, and P-126 are isochrones for 11,800 years, 12,000 years, and 24,000 years.

Schott did not state that distortion had been considered. The diagrams of the cores given by Schott (1935, Beiläge I, II, III) show that the top of the layer which is free from *G. menardii* is close to the bottom of many of his cores. Recent measurements of the distortion in core samples indicate that the length of the core above this boundary must be considerably less than the depth of sediment sampled. Therefore, Schott's average values, expressed in cm. per 1000 years, are probably somewhat low. His values, however, are based on the assumption that the *G. menardii* began to accumulate 20,000 years ago in the postglacial time of the equatorial Atlantic. The uncertainty of this assumption is possibly as great as the error introduced by failure to correct for the distortion.

NORTH ATLANTIC SEDIMENTS

In Figure 4, the rate of deposition in cm. per 1000 years is plotted against time. The various zones in the three cores from the North Atlantic illustrated in Figure 2 are replotted against time instead of core depth, a few core depths being inserted for orientation. Isochrones connect points of correlated zones to illustrate the degree of contemporaneity. The time scales, from left to right, are in the ratios 1 : 2 : 6.25 : 25 : 25. Rates of deposition for the sediments represented by core P-124(3) are very high, either as based on the present measurements or on any previous estimates. This may be expected from the position of this core at the foot of the Grand Banks slope and in the Labrador Current. The rapid deposition of the sediments in core P-130(9) is similarly attributed to its relation to the Faraday Hills of the Mid-Atlantic Ridge.

Both cores give similar indications of the changes in the rate of sedimentation during the past 12,000 years. The rate of deposition in late postglacial time has increased and is unusually high at present. This is demonstrated in Figure 4 as applying to such widely separated areas as the Caribbean Sea, the Pacific Ocean, and the North Atlantic.

There seems to be some evidence (Fig. 4) that the deposition of the last glacial marine sediment was more rapid than that of the immediately ensuing warm-water deposit. On the other hand, deposition is more rapid after a prolonged period of warm surface water than during the periods of glacial marine deposit. The rates in the top of core P-126(5) are omitted in Figure 4 because, for any reasonable accuracy, the rapid change in the radium content with depth near the top requires the determination of the radium at closer intervals than is feasible. A high rate at present is indicated, however, as in the other North Atlantic cores. The lower part of the zone of intermediate character in core P-126(5), between the upper two zones of glacial marine deposit, is characterized by a low rate of sedimentation which is understandable, because the depo-

sition of detrital material diminished and floor building was more dependent upon organisms existing at relatively low water temperatures. Similarly, following the last deposition of glacial marine sediments, there are indications that considerable time was necessary, after the prolonged period of cold surface water, to re-establish the normal abundance of organic life. This period may be measured in terms of some thousands of years. Most of the small fluctuations in Figure 4 are within the experimental error; in fact, the accuracy does not warrant tabulating the data in Figure 4.

The zones of anomalous deposits in core P-126(5) show abnormal rates of sedimentation. The rates as drawn in Figure 4 in the region of these zones were calculated after their thicknesses were subtracted, and they indicate that there was an increased rate of deposition, exclusive of the anomalous deposits, just prior to the last glacial marine sedimentation. The rates for the upper and lower anomalous zones, respectively, are 35 (39) and 15 (18) cm. per 1000 years. As previously mentioned, these high figures might well indicate almost instantaneous precipitation, because they are based on intervals of time so short as to be extremely dependent upon the drawing of the experimental curve. In this respect, the results support the theory of submarine slumping (Bradley *et al.*, 1940).

The absolute values for the rates of deposition in cores P-124(3) and P-130(9) must be considered as representative of abnormal conditions. Those for core P-126(5) probably indicate the order of magnitude for the general region west of the Mid-Atlantic Ridge.

CARIBBEAN SEA SEDIMENTS

The rate of deposition of the Globigerina ooze in the Cayman Trough (Caribbean Sea, core P-137) is roughly constant between 10,000 and 300,000 years at 0.5 to 1.0 cm. per 1000 years, but it increases during postglacial time to 4 cm. per 1000 years at present. The rates for this Globigerina ooze as given by the dashed curve are probably more accurate than those given by the solid curve, particularly for this core. If this is correct, the glacial ages are characterized by somewhat lower rates of deposition. This is contrary to the estimates of Schott (1939, p. 413) for the last Wisconsin age, but examination of core P-137 shows no increase in the amount of detrital material more than sufficient to offset a decrease in the abundance of organisms during these ages, and there is no contradiction in these opposing results. This increase in the amount of detrital material was Schott's explanation for the increased rates in the last ice age.

RED CLAY FROM THE PACIFIC OCEAN

The rates of deposition of Red clay (P-259) are given for the upper 160 cm. only. As has been pointed out, the method becomes inaccurate beyond 300,000 years. This core was secured by a different technique (Emery and Dietz, 1939; 1941), and hence the application of the distortion data, obtained with the authors' coring apparatus, is doubtful. There are indications that the rate of deposition has increased in recent times. Certainly the rate of sedimentation of Red clay, as exemplified by this core, is lower than that of the other types of ocean sediments which have been studied, and in this respect the results agree with previous estimates. The mean rate for the past 300,000 years is 0.5 (0.6) cm. per 1000 years, which may be compared with Twenhofel's (1929) estimate of 0.35 cm. of inorganic material deposited on the sea bottom, remote from coasts, in 1000 years. Schott's (1935) estimate for Red clay is <0.86 cm. per 1000 years, with a maximum observed rate of 1.33 cm. and a minimum of <0.5 cm. The rate of deposition of this Red clay, as shown in Figure 4, is 0.25 cm. per 1000 years, 300,000 years ago, or 0.35 cm. if the distortion data be applied. The remaining 86 cm. of core below this point must represent at least 86 cm. of sediment (246 less 160 cm.) and possibly 175 cm., if the distortion data can be applied. Assuming the rates in each case to be those just quoted for the 300,000 year horizon, one should add 350,000 or 500,000 years for the lower portion of the core, to which the present method of dating is inapplicable.

Emery and Dietz (1941) are of the opinion that the ratio of core length to penetration is constant at 0.73 in this Red-clay core. The rates of sedimentation will then be given by the solid line in Figure 4 (P-259) multiplied by the factor 1/0.73. By comparison with the rates of deposition calculated with the distortion data (dashed curve, Fig. 4) these rates are higher near the top of the core but become nearly the same beyond 100,000 years. The mean rate for the past 300,000 years is 0.7 cm. per 1000 years, and the time required to deposit the lower portion of the core remains the same as if no distortion took place—namely, 350,000 years—giving a total time of 650,000 years for the sedimentation of the whole core.

COMPACTION

Apart from the distortion introduced by sampling, there may be compaction of the ocean sediments as they lie in the bottom. The apparently high rate of deposition today may be due in part to compaction, as the sediments become buried. The answer to this problem is given by expressing the results in Figure 4 in terms of milligrams of water-free matter deposited per square centimeter per year. The conversion of

units requires a knowledge of the weight of dried material in a unit volume of the original sediment. The measurements of this quantity have been described elsewhere (Piggot and Urry, 1942; Urry and Piggot, 1942).

TABLE 2.—*Rates of deposition of ocean sediments in milligrams of water-free matter per square centimeter per year*

Time interval in the past (years)	Zone[a]	Mean rate of deposition in milligrams per square centimeter per year	
		A[b]	B[b]
		P-124 (3) N. Atlantic	
0–1,000	WS	39	41
1,000–1,500	WS	38	43
1,500–2,000	WS	32	36
2,000–4,000	WS	22	27
4,000–10,000	WS	15	25
10,000–12,000	WS	11	32
		P-130 (9) N. Atlantic	
0–1,000	WS	21	21
1,000–3,000	WS	13	15
3,000–4,000	WS	8	10
4,000–12,000	WS	6	7
12,000–20,000	W-G	7	11
20,000–24,000	G	7	19
		P-126 (5) N. Atlantic	
0–8,000	WS	—	—
8,000–12,000	WS	0.9	1.0
12,000–15,000	W	2.2	2.3
15,000–24,000	G	4.3	4.7
24,000–28,000	HW	7.9[c]	9.1[c]
28,000–42,000	HW	2.2	2.8
42,000–51,000	G-W-G	4.2	5.9
51,000–62,000	WS-W-G	3.6	5.8
62,000–73,000	W	3.4	8.4
		P-137 Globigerina ooze	
0–2,500		5.0	5.1
2,500–4,500		3.0	3.2
4,500–8,500		1.5	1.5
8,500–200,000		0.7	0.9
200,000–300,000		0.4	1.2
		P-259 Red Clay	
0–20,000		0.7	0.7
20,000–60,000		0.7	0.7
60,000–160,000		0.4	0.45
160,000–300,000		0.2	0.25

a. This column gives the zones illustrated in Figures 2 and 4 as follows: W, warm-water deposits; S, volcanic shards present; HW, the warm-water? zone of intermediate character; G, glacial marine deposits.

b. Column A gives the rate of deposition with the assumption that the core length represents the sediment sampled. Column B gives the rate of deposition with the assumption that the relations between core length and the depth of sediment sampled are those found for the varved clays of Connecticut. The true rate of deposition is probably closer to the values given in column B.

c. These rates do not include the number of centimeters and hence the mass of material in the anomalous zones. (*See* Figure 2.)

DEPOSITION IN TERMS OF MASS

Table 2 summarizes the rates of deposition expressed in milligrams per square centimeter per year. As in Figure 4, two sets of results must be given for the limiting cases of distortion. The relative fluctuations of the rates of deposition are qualitatively similar to those presented in Figure 4, and, in particular, the high rate of deposition at present, as measured in centimeters per 1000 years, is due to an increase in the mass of material precipitated in recent times.

SUMMARY

The ocean floor is covered by deposits of unknown thickness which, for the most part, are stratigraphically undisturbed. Representative samples of several meters of this sediment can now be obtained from below as much as 6000 meters of water. Such core samples provide abundant lithological, chemical, and biological evidence for all the major events occurring within the time that they represent, which in some cases is nearly all of the Pleistocene. Longer cores may eventually penetrate the Pliocene. If these events can be dated, a considerable advance is possible in the chronology of the Pleistocene epoch, for the events occurring not only in the oceans, but also on the continents.

Radium measurements throughout the length of the cores were undertaken in the Geophysical Laboratory, primarily to solve the problem of the high radium concentrations in the surface of some ocean sediments (Piggot, 1933). The solution of this problem provided the basis for a method of measuring time in ocean-core samples. If there are not certain ratios between the numbers of the various types of radioactive atoms incorporated in the ocean sediments, the concentrations of these radioelements will change, with time, in such a manner as to establish these ratios. By following the change of concentration of an appropriate element, this property of radioactivity provides a method of measuring intervals of time within the past 300,000 years. It does not involve the accumulation of either of the stable products of radioactive disintegration—*i.e.*, helium or lead—but requires only a knowledge of the relative concentrations of a radioelement at different depths in the sediment.

The radioelements are introduced into the sediments in amounts differing from the "equilibrium" ratio mentioned above. This accounts for the high radium content of certain deep ocean deposits, compared to similar concentrations in continental formations (Piggot and Urry, 1941). The dates in a core once having been established, a calculation of the rate of deposition should be simple, but, while there is little doubt that the core represents everything that it penetrates, there is no doubt that the length of a particular portion of core is not always the same as the sedi-

ment represented by that sample, particularly near the bottom of the core. This introduces uncertainty into the estimates of rates of deposition derived from previous studies of cores and impairs, to some extent, the present study even though measurements were made to determine the distortion with the same core sampler used to obtain four of the five cores discussed here.

Cores of the sediments of the North Atlantic were studied in great detail by members of the U. S. Geological Survey and their collaborators. Dates are given here for the geological events, in the North Atlantic, discovered in their studies.

Time intervals have been calculated for one core of Globigerina ooze, underlying the Caribbean Sea, which may be applied to Cushman's detailed survey of the Foraminifera in this core.

The radium in a core of Red clay from the Pacific Ocean was determined with the object of throwing some light on the high concentration of radium in the surface of Red clays in general and in order to study the rate of deposition of Red clay.

The rates of deposition in the past, as well as at present, are given for all the core samples studied. The rates given for the upper portions of the cores are probably nearly correct, but one should be cautious in drawing conclusions from changes in the rate in the lower part of a core. However, there is little doubt that deposition is as rapid today as at any time during the period represented by the cores studied here, if not more rapid.

NOTES ON THE CORES OF OCEAN-BOTTOM SEDIMENTS

P–124 Collected on board the cable ship LORD KELVIN May 1936. Station No. 3. Lat. 46° 03′ N., Long. 43° 23′ W. Depth of water 4710 meters. Length of core 285 cm.

P–126 Collected on board the cable ship LORD KELVIN May 1936. Station No. 5. Lat. 48° 38′ N., Long. 36° 01′ W. Depth of water 4760 meters. Length of core 282 cm.

P–130 Collected on board the cable ship LORD KELVIN May 1936. Station No. 9. Lat. 49° 40′ N., Long. 28° 29′ W. Depth of water 3750 meters. Length of core 276 cm.

P–137 Collected on board the research ketch ATLANTIS February 1937. Lat. 19° 14′ N., Long. 80° 20′ W. Depth of water 4893 meters. Length of core 190 cm.

P–259 Collected on board the motor vessel E. W. SCRIPPS by F. P. Shepard and R. Revelle with their core sampler. Core No. FPS 186, August 1938. Lat. 30° 41′ N., Long. 121° 46′ W. Depth of water 4080 meters. Length of core 246 cm.

REFERENCES CITED

Antevs, Ernst (1928) *The last glaciation,* Am. Geog. Soc., Res. Ser. No. 17, New York.

Bradley, W. H., *et al.* (1940) *Geology and biology of North Atlantic deep-sea cores between Newfoundland and Ireland. Foreword* by C. S. Piggot. *General introduction* by W. H. Bradley. Part 1, *Lithology and geologic interpretations* by M. N. Bramlette and W. H. Bradley. Part 2, *Foraminifera* by J. A. Cushman and L. G. Henbest, U. S. Geol. Survey, Prof. Paper 196-A, p. I–XV + 1–56.

Chamberlin, T. C., and Salisbury, R. D. (1906) *Geology,* vol. 3, Henry Holt and Co., p. 392–394.

Cushman, J. A. (1941) *A study of the Foraminifera contained in cores from the Bartlett Deep,* Am. Jour. Sci., vol. 239, p. 128–147.

Emery, K. O., and Dietz, R. S. (1939) *Telescoping in marine cores* (Abstract), Geol. Soc. Am., Bull., vol. 50, p. 1908.

——— — ——— (1941) *Gravity coring instrument and mechanics of sediment coring,* Geol. Soc. Am., Bull., vol. 52, p. 1685.

Kay, G. F. (1931) *Classification and duration of the Pleistocene period,* Geol. Soc. Am., Bull., vol. 42, p. 425–466.

Lane, A. C. (1924-1941) *Annual Reports of the Committee on the measurement of geologic time,* Nat. Res. Council, Washington, D. C., private printing.

——— (1937) *Annual Report of the Committee on the measurement of geologic time,* Nat. Res. Council, Washington, D. C., private printing, p. 4.

Lohmann, H. (1909) *Plankton-Ablagerungen am Boden der Tiefsee,* Naturwiss. Ver. Schleswig-Holstein Schr., Band 14, p. 399–403.

Murray, J., and Peake, R. (1904) *On recent contributions to our knowledge of the floor of the North Atlantic Ocean,* Royal Geog. Soc., London, p. 1–35.

Penck, A., and Brückner, E. (1901–1909) *Die Alpen in Eiszeitalter,* 3 vols., Leipzig.

Piggot, C. S. (1941) *Factors involved in submarine core sampling,* Geol. Soc. Am., Bull., vol. 52, p. 1513–1523.

———, and Urry, W. D. (1939) *The radium content of an ocean-bottom core,* Washington Acad. Sci., Jour., vol. 29, p. 405–410.

——— — ——— (1941) *Radioactivity of ocean sediments. III. Radioactive relations in ocean water and bottom sediment,* Am. Jour. Sci., vol. 239, p. 81–91.

——— — ——— (1942) *Radioactivity of ocean sediments. IV. The radium content of sediments of the Cayman Trough,* Am. Jour. Sci., vol. 240, p. 1–12.

Schlundt, H. (1934) *The radium content of some travertine deposits at Mammoth Hot Springs, Yellowstone Park as an index of their age* (Abstract), Exhibit E, Annual Report of the Committee on the measurement of geologic time, Nat. Res. Council, Washington, D. C., private printing, p. 34–38.

Schott, W. (1933) *Die jüngste Vergangenheit des äquatorialen Atlantischen Ozeans auf Grund von Untersuchungen an Bodenproben der Meteor-Expedition,* Naturf. Gesell. Rostock Sitzungsber. Abh. Ser. 3, Band 4, p. 48–57.

——— (1935) *Die Foraminiferen in dem äquatorialen Teil des Atlantischen Ozeans,* Wiss. Ergeb. Deutschen Atlantischen Exped. Meteor 1925–1927, Band 3, Teil 3, Lief. 1B, p. 43–134.

——— (1939) *Rate of sedimentation of recent deep-sea sediments, Symposium, Recent marine sediments,* Part 6, Publ. by Am. Assoc. Petrol. Geol., Thomas Murby and Co., p. 409–415.

Schuchert, C. (1931) *Geochronology or the age of the earth on the basis of sediments and life,* in *Physics of the earth, IV, The age of the earth, Part I,* Nat. Res. Council, Bull. 80, Washington, D. C., p. 10–64.

Twenhofel, W. H. (1929) *Magnitude of the sediments beneath the deep-sea,* Geol. Soc. Am., Bull., vol. 40, p. 385–402.

Urry, W. D. (1941) *The radioactive determination of small amounts of uranium,* Am. Jour. Sci., vol. 239, p. 191–203.

——— (1942) *The radio-elements in non-equilibrium systems,* Am. Jour. Sci., vol. 240, p. 426–436.

———, and Piggot, C. S. (1942) *Radioactivity of ocean sediments, V. Concentrations of the radio-elements and their significance in Red clay,* Am. Jour. Sci., vol. 240, p. 93.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 53, PP. 1211-1230, 2 PLS., 2 FIGS. AUGUST 1, 1942

GLACIATION OF SHICKSHOCK MOUNTAINS, GASPE PENINSULA

BY RICHARD FOSTER FLINT, MAX DEMOREST, AND A. L. WASHBURN

CONTENTS

Page

Abstract 1211
Introduction 1212
Acknowledgments 1213
Historical summary 1213
Shickshock Mountains 1215
General description 1215
Mount Albert 1215
Lithology 1215
Topography 1215
Evidence of glaciation 1217
Erratic stones 1217
Glaciated surfaces 1219
Character of Mount Albert glaciation 1219
Tabletop Mountain 1220
Lithology 1220
Topography 1220
Evidence of glaciation 1221
Glacial striae 1221
Erratic stones 1221
Broad glaciated surfaces 1222
Close relation of mantle to underlying rock 1223
Summary 1223
Case for nonglaciation 1224
Date of glaciation 1226
Source of glacier 1227
References cited 1229

ILLUSTRATIONS

Figure Page

1. Index map of Gaspe Peninsula 1214
2. Map of Mount Albert-Tabletop area 1216

Plate Facing page

1. Summits of Mount Albert and Tabletop 1218
2. Views on Tabletop 1219

ABSTRACT

The Shickshock Mountains constitute one of the highland areas that have been considered by Coleman on geologic evidence, and by Fernald on botanical and

geologic evidence, to have been nunataks during the maximum of Wisconsin glaciation.

Recent field studies have been made on the two highest parts of the Shickshocks—the broad plateaulike masses known as Mount Albert (summit altitude 3775 feet) and Tabletop Mountain (4230 feet). Glacial erratics possibly derived from pre-Cambrian rocks from north of the St. Lawrence were found as high as 3760 feet on Mount Albert. On Tabletop Mountain, striated surfaces were found as high as 3500 feet, and glacial erratics (of local origin) as high as 3700 to 3800 feet. Detailed search failed to produce any direct evidence of glaciation through the highest 400-500 feet. Throughout this distance evidences of any possible glaciated surfaces are obscured by mantles of locally derived felsenmeer that presumably originated in postglacial time. In addition, the composition of the bedrock is so heterogeneous that several types of erratic stones might be present without having been recognized. These circumstances do not demonstrate glaciation of the highest parts of Tabletop Mountain; yet there is no geologic evidence inconsistent with the possibility that the entire Shickshock highland was overtopped by Wisconsin ice. No geologic evidence was found suggesting that any portion of the area existed as a nunatak during the Wisconsin maximum.

During the waning stages of glaciation, and possibly earlier, Tabletop was the site of radial outflow from a local ice cap.

INTRODUCTION

The glacial history of the Shickshock Mountains of the Gaspe Peninsula has been variously interpreted, the chief point at issue being whether or not the highest summits existed as nunataks during the maximum of Wisconsin glaciation. Coleman (1922, p. 14, 37) on the basis of geologic evidence and Fernald (1925, p. 295-303) on the basis of both botanical and geologic evidence have favored the nunatak hypothesis. Mailhiot (1919, p. 146) and Alcock (1935, p. 125-126), on the other hand, expressed the opinion that some of the highest summits have been glaciated. As a result of field investigations in August 1941, the present authors are convinced that the existence of nunataks has not been proved. However, they could find no proof that the highest summits were glaciated. It is nevertheless demonstrated that at some time ice reached to within at least 500 feet of the highest summit (Mt. Jacques Cartier) and was at least 1200 feet higher than believed by Coleman. It is thought that this highest ice was of Wisconsin age; it may have been part of the great Labradorian ice sheet or it may have been part of a local ice cap centered over the Tabletop plateau on which Mt. Jacques Cartier is situated.

This study and others in similar highland areas such as the Torngat Mountains in Northern Labrador (Odell, 1938, p. 204-207), Mt. Katahdin in Maine (Tarr, 1900; R. P. Goldthwait, unpublished communication), Mt. Washington (R. P. Goldthwait, 1940, p. 17-19), the Long Range of Newfoundland (MacClintock and Twenhofel, 1940), and the Adirondack Mountains (J. P. Marble, R. Balk, unpublished communications) show that there is no apparent geologic evidence that highland areas in eastern North America existed as nunataks during the maximum of Wisconsin glaciation.

ACKNOWLEDGMENTS

The field work was financed by a grant from the Dana Research Fund of Yale University. Thanks are due Dr. F. J. Alcock of the Geological Survey of Canada, Ottawa, and Dr. I. W. Jones of the Quebec Bureau of Mines, Quebec City, for helpful discussion prior to the field work and for examining glacial erratics and giving opinions on their probable sources. The Quebec Department of Fish and Game kindly authorized entrance into the Gaspesian Park, although this area is at present officially closed. The many kindnesses and courtesies of park and fire-lookout personnel, especially Samuel Coté, Gustave Dupuis, and Paul Pelletier, are gratefully acknowledged. Charles-Edouard Levesque of Ste. Anne des Monts served very efficiently as guide to the party. Mrs. A. L. Washburn, who accompanied the party, aided the work in various ways.

HISTORICAL SUMMARY

Apparently the first geologist to mention the glacial features of the summit plateaus of the Shickshocks was Mailhiot (1919, p. 146) who held that the summit of Mt. Albert had been glaciated. Coleman (1922, p. 12-14, 31-39), after visiting the summits in 1918 and 1919, concluded that the Tabletop and Mt. Albert plateaus had not been crossed by glacier ice and that Tabletop had not even been covered by a local ice cap. However, he recognized that Mt. Albert might have been a center of radial glaciation. Also, he showed that the Labrador ice sheet crossed Gaspe in the region of the low pass (altitude 750 feet) at the head of the Matapedia River, west of the Shickshocks.

In 1921-1924 Alcock (1926, p. 23-25) studied the geology of the Mt. Albert area and also visited Tabletop and inferred, as Coleman had done, that the summits had not been glaciated. However, after several years' further field study of Gaspe, he (Alcock, 1935, p. 125-126) changed this opinion and concluded that the summit areas probably had been covered by the Labradorian ice sheet.

In 1932 Jones (1933, p. 28) investigated the geology of Tabletop and found no evidence of glaciation at altitudes higher than the heads of the cirques recognized by Coleman.

A botanical party led by M. L. Fernald visited Tabletop and Mt. Albert in 1904. Much later, Fernald (1925, p. 295-303) correlated the distribution of plants in this region with the distribution of glaciated and non-glaciated areas as inferred by Coleman (1922) and later by Alcock (1926).

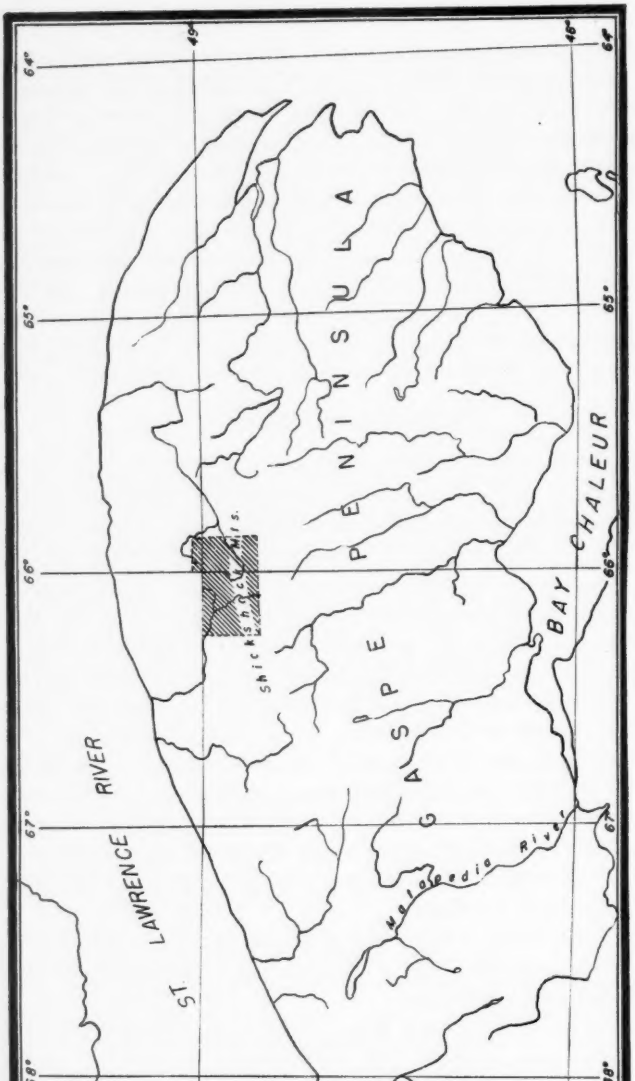


FIGURE 1.—*Index map of Gaspe Peninsula*
Area of Figure 2 is shown by shading.

SHICKSHOCK MOUNTAINS

GENERAL DESCRIPTION[1]

The Shickshock Mountains, lying north of the central part of the Gaspe Peninsula, constitute its highest part. The Shickshock axis is essentially a continuation, via the Notre Dame Mountains of Southeastern Quebec, of the Green Mountains of Vermont. Trending east-northeast, the Shickshocks constitute a belt having a maximum width of 15 miles and lying no more than 25 miles south of the St. Lawrence. They consist of deformed Paleozoic sedimentary and volcanic rocks with some large intrusive masses, the whole belt being slightly convex toward the north.

The areas underlain by resistant rocks—the volcanics and intrusives—have broad summits with nearly flat to very irregular surfaces. Alcock (1935, p. 109) relates these areas to an erosion surface that has been warped up and isolated through reduction of the surrounding weak-rock areas during a later cycle of erosion.

The highest parts of the Shickshock Mountains, lying close to the intersection of the 49th Parallel and the 66th Meridian, consist of Tabletop Mountain, whose highest summit reaches 4230 feet, and Mt. Albert, with an extreme altitude of 3775 feet (Figs. 1, 2). Both masses are extensive and are developed on resistant intrusive rocks.

The Shickshock region is densely wooded, chiefly with spruce, fir, and birch, but above altitudes of 2800 to 3300 feet trees (mostly stunted spruce) occur only in patches separated by extensive areas of mosses, grasses, and bare rock.

MOUNT ALBERT

LITHOLOGY

The higher portions of Mount Albert (Fig. 2) consist of serpentinized peridotite weathered buff, encircled by dark amphibolite. (*Cf.* Coleman, 1922, p. 31; Alcock, 1926, p. 34-38.) The serpentinized peridotite characterizes the south peak and all of the plateau except the narrow ridge on the north side, which is composed of the amphibolite.

TOPOGRAPHY

The most striking topographic feature of Mount Albert is the plateau-like character of the higher portions (Pl. 1, fig. 1). The name "tabletop" is much more descriptive of Mount Albert than it is of Tabletop Mountain. The Mount Albert plateau has roughly the shape of a thick-set, irregular letter C with the concavity opening eastward. From north to

[1] An excellent discussion of the physical features of the Gaspe Peninsula and their origin is given by Alcock (1935, p. 109-119).

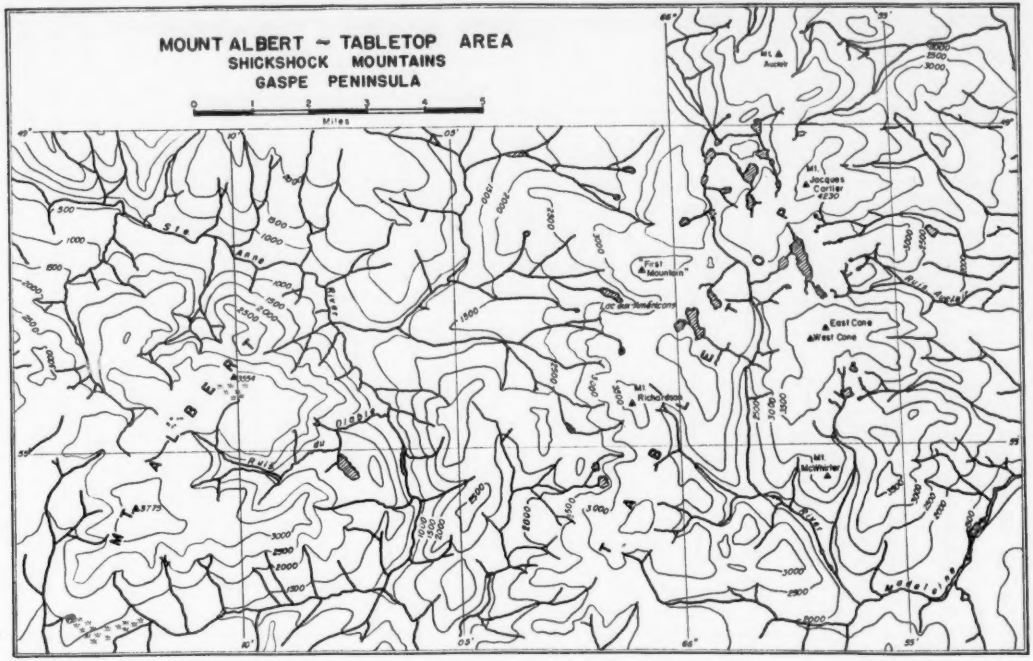


FIGURE 2.—*Map of Mount Albert-Tabletop area*

Compiled from maps of Geological Survey of Canada and Quebec Bureau of Mines. (*See* Alcock, 1926, and Jones, 1933.) Because of differences between the maps the present compilation includes adjustments along the 66th Meridian.

south the plateau is about 3 miles long. In an east-west direction the two arms of the C are about 2 and 3½ miles long, the southern arm being the longer. The plateau is bounded on all sides by rather abrupt slopes, some of which descend steeply to the base of the mountain. Cirquelike valley heads indent the plateau in several places, a notable one being on the east side at the head of Rivière du Diable. This valley head is chiefly responsible for the C-like shape of the plateau.

The highest point of Mount Albert is the south peak, with an altitude of 3775 feet. The north peak, forming the highest point on the amphibolite ridge about 3 miles away, has an altitude of 3554 feet and is the more abrupt. From the base of the north peak and from the crest of the south peak, the surface slopes gently and evenly down to about 3300 feet at the lowest and narrowest portion of the plateau, approximately midway between the two peaks. The maximum differential relief on the plateau is thus about 500 feet. A number of small lakes constitute a striking feature of the low, narrow portion. Alpine types of vegetation dominate some areas of the plateau. Other areas are characterized by bare-rock knobs or by felsenmeer—continuous sheets of fragments of frost-disrupted local rocks in process of slow transport down the slopes. Stone-polygons are well developed in places.

EVIDENCE OF GLACIATION

Erratic stones.—Erratics are far from common on the Mount Albert plateau. Possible explanations of their scarcity here and on the other highland areas are discussed in another section of this report. Mailhiot (1919, p. 146) was the first to record the presence of erratics. He says: "On the summit plateau are found a few erratic blocks of granite and hornblende schist which certainly represent remnants of ancient glacial moraines." According to Coleman (1922, p. 31), "The mountain is really a steeply walled tableland covered with weathered masses of serpentine except for a ridge of garnetiferous hornblende schist which rises a little higher than the rest and has rolled fragments down on the serpentine." Coleman's explanation of the occurrence of amphibolite on serpentinized peridotite may be correct for the immediate vicinity of the ridge, but this explanation does not seem possible for an amphibolite erratic which one of the present authors found within 100 feet of the summit of the south peak and thus about 100 feet higher than the summit of the amphibolite ridge 3 miles away.

Mailhiot (1919, p. 146) mentions granites on the plateau, *i.e.*, at altitudes of more than 3300 feet. Coleman (1922, p. 32) mentions granite boulders up to about 1500 feet only. According to an unpublished communication from F. J. Alcock, a gold-quartz boulder of probable Labra-

dorian origin was recently found at an altitude of about 1550 feet, about 6 miles south of the summit of Mount Albert.[2] Alcock (1926, p. 24) found one 10 feet in diameter at an altitude of 2500 feet and states that two others occur at a similar altitude. Alcock also records a conglomerate boulder about 20 feet long at an altitude of about 3200 feet.

The present investigation disclosed erratics at still higher altitudes. Quartzites were found on the plateau at altitudes of over 3300 feet; a granite-gneiss cobble was collected within 50 feet of the summit of the south peak, and a granite pebble within 15 feet of the same summit. Stones of serpentinized peridotite were observed within 15 feet of the summit of the north peak. According to Alcock (unpublished communication) the highest erratic and several others found at lower altitudes (biotitic quartzite or paragneiss at about 3450 feet; garnetiferous sericite schist at about 3100 feet) are probably of pre-Cambrian origin from across the St. Lawrence and hence may have been transported by Labradorian ice. However, as pointed out by Alcock and emphasized by Jones (unpublished communication), the erratics are not sufficiently distinctive to prove such an origin.

That these erratics could not have been deposited by streams on the preglacial erosion surface represented by the plateaus of Mount Albert and Tabletop and other upland areas (Alcock, 1935, p. 109) is demonstrated by the following evidence: (1) The granitic rocks observed on the south peak are firm and but little weathered and could hardly have been subjected to weathering throughout Pleistocene time; in fact their degree of weathering is consistent with a Wisconsin age; (2) the quartzites and conglomerate rocks must have been derived from lower places; (3) several of the erratics are too large to have been transported by streams on an erosion surface of peneplanelike character; (4) the serpentinized peridotite boulders on the north peak can hardly be regarded as derived from the higher serpentinized peridotite area by mass-wasting or fluvial transport. The present topography would not permit this, and, if topographic reversal has occurred in the vicinity of the north peak since their deposition, the serpentinized peridotite erratics would have had to survive while a considerable prism of bedrock of the same lithology was completely removed. Thus, with the possible exception of the amphibolite stones lying on serpentinized peridotite at the base of the north peak, all the erratics must have been glacially transported. It therefore follows that ice covered Mount Albert up to within at least 15 feet of the top of the highest peak.

[2] The boulder, 3½ feet in diameter, was found by Mr. Beidelman, of the Federal Zinc and Lead Company, about 100 feet above Brandy Brook on its northwest side, near the northwest corner of mining claim 1563. (Cf. Alcock, 1928, Map 2129, 210A.)

FIGURE 1. MOUNT ALBERT PLATEAU
View from north peak toward south peak.

FIGURE 2. PLATEAULIKE SURFACE OF TABLETOP
View from near "First Mountain" northeast toward Mount Jacques Cartier.

SUMMITS OF MOUNT ALBERT AND TABLETOP

FIGURE 1. LARGE GRANITE ERRATIC On summit of "First Mountain".

FIGURE 2. ROCK-BASIN LAKES Steep slope in foreground is head of Lac aux Américains cirque.

FIGURE 3. FELSENMEER WITH STONE-POLYGONS Summit of Mount Jacques Cartier.

VIEWS ON TABLETOP

Glaciated surfaces.—A number of irregular rock-basin lakes, several yards to several tens of yards in diameter, occur in the low, narrow portion of the plateau at about 3300 feet. Being in serpentinized peridotite, the basins cannot be of solutional origin, and the writers know of no other process except glaciation which would form basins of this character. Moreover, the fact that even very small rock-basin lakes are not filled with sediment argues for recency of glaciation.

No other definite products of glaciation were noted on the Mount Albert plateau. In several places there is a suggestion of stoss-and-lee topography which could be interpreted to indicate ice movement from the northwest across the plateau (*cf.* Mailhiot, 1919, p. 146); a tendency for the lakes to be elongate in a northwest-southeast direction could be taken as support for such an interpretation. However, as jointing in various directions is well developed in the serpentinized peridotite, it is possible that such directional features in the topography are structurally controlled.

The fact that no striae or other glacial markings were observed by Coleman, Alcock, and the present writers is not to be wondered at, for the surfaces of the bare-rock areas are so much weathered that they could not be expected to retain such features.

CHARACTER OF MOUNT ALBERT GLACIATION

Coleman (1922, p. 31) grants that ". . . ice may have moved outwards from it [Mount Albert]." According to Alcock (1926, p. 24),

> "Local glaciers occupied the flanks of Mount Albert during the Pleistocene. On the west and south sides of the mountain are a number of amphitheatres which are undoubtedly cirques formed by small glaciers. A rather striking amphitheatre lies about a mile west of the north geodetic station. It is well defined, with steep walls 500 feet high and a flat bottom in which lies a small lake. Other steep-walled cirques to the south have a depth of 1,000 feet."

Later Alcock (1935, p. 124-126), on the basis of indirect evidence, suggested that, in addition to being the site of local glaciation, the high Shickshocks, including Mount Albert, were crossed by Labradorian ice, thus abandoning his earlier (1926) view that this had not occurred.

Observations of the present authors prove that the summit plateau was also glaciated. The fact that rocks foreign to the plateau occur among the erratics shows that the ice which covered the plateau originated elsewhere. However, the lithology of the erratics does not prove whence the ice came. As previously pointed out some of these erratics may have originated in localities north of the St. Lawrence River, in which case they were brought to the Mount Albert region by Labradorian ice. However, even if such Labradorian origin were demonstrable, there would still be no proof that Labradorian ice ever covered the summit of Mount

Albert. This is because there is a possibility that Labradorian ice may have deposited erratics at a low altitude in valleys between Mount Albert and Tabletop and that subsequent redistribution by radial outflow from an ice cap on the Tabletop plateau brought them to the summit area of Mount Albert.

Thus it appears that Mount Albert has been the site of two kinds of glaciation: (1) General glaciation that covered Mount Albert. This was either continental (Labradorian), or local (originating on Tabletop), or both; (2) strictly local glaciation; certainly cirque glaciation, perhaps preceded by a small Mount Albert ice cap.

TABLETOP MOUNTAIN

LITHOLOGY

Tabletop Mountain (Pl. 1) consists of a granitic intrusive about 10 miles long by 5 miles wide and of irregular form. The rock varieties seen by the writers consist of granite, syenite, granodiorite, diorite, granite porphyry, felsite, and pegmatite. These have been described by Jones (1933, p. 24-25) who believes that they are differentiates of a single magma. Surrounding the intrusive is a rather narrow zone of somewhat altered rocks, consisting chiefly of slates altered to hornfels and calcareous sediments altered to lime-silicate rocks.

TOPOGRAPHY

Tabletop Mountain is a kind of dissected plateau, although the surface is not what is ordinarily thought of as plateaulike, inasmuch as the maximum relief of this small area is nearly 1000 feet, exclusive of the capacious valley of the Madeleine River that is incised into the southern half of its central part.

The outer flanks of the plateau are very steep slopes, in places 2000 feet high. These are indented by deep valleys with cirquelike heads, the altitudes of the floors averaging a little more than 2500 feet. The plateau has a high irregular rim held up by resistant rocks: chiefly metamorphics on the west side and chiefly granites and related rocks on the east side. The main area inclosed within this rim is several hundred feet lower than the rim itself and is occupied by a number of lakes.

The domelike peaks that together constitute the rim are nearly bare of vegetation and are extensively covered with felsenmeer. In places this superficial debris is organized into stone-polygons and stone-stripes. The highest of these peaks is Mt. Jacques Cartier with an altitude of 4230 feet (Pl. 1, fig. 2); others are Mt. Auclair, Mt. Richardson, Mt. McWhirter, and the broad ridge that bears East Cone and West Cone.

EVIDENCE OF GLACIATION

Glacial striae.—Although Coleman (1922, p. 14) found no striae at altitudes greater than 2500 feet, the writers found striae at two localities at much higher altitudes. One locality (Fig. 1) is at an altitude of about 2900 feet on the trail ascending the west flank of Tabletop Mountain from Lac aux Américains. The striae occur on a smooth boss of lime-silicate rock that has a surface sloping in a direction S. 12° W. at about 50° from the horizontal. On this surface the striae trend S. 83° W. and slope down toward the southwest at 18° from the horizontal. One hundred feet higher on this same trail is another boss of similar rock with striae trending S. 82° E.

In the writers' opinion these two occurrences of striae are of glacial origin and therefore demonstrate glaciation at this place. Unfortunately the direction of ice flow cannot be surely inferred from them so that no discrimination could be made between the possibility of ice encroaching upon the Shickshock highland from the region north of the St. Lawrence and the possibility of ice flowing radially outward from the Shickshock highland itself.

The other locality at which striae occur is on the north flank of the peak called by Coleman "First Mountain," nearly a mile northeast of the first locality and at an altitude of about 3500 feet. The rock here is hornfelslike. One boss has a face sloping north at about 45° from the horizontal; on this face are striae believed to be glacial, trending obliquely to it, but no features were observed that would indicate the direction of ice flow.

All that can surely be inferred from the observed striae is that glaciation on the west side of Tabletop Mountain up to an altitude of at least 3500 feet is demonstrated. Although a search was made for other striae, none was found. This is explained at least in part by the fact that the low central area of the plateau is covered with vegetation and the barren eastern part consists largely of granite whose surfaces, roughened by weathering at these altitudes, are not capable of preserving striae. In any case striae on the east side of Tabletop would not discriminate between Labradorian ice and local ice, for either would have flowed in the same direction, namely, southeast.

Erratic stones.—On the west rim of Tabletop, stones and boulders of Tabletop granite rest on the flanks and summit of "First Mountain," a peak consisting entirely of metamorphic rocks. These stones mentioned by both Coleman (1922, p. 37) and Alcock (1926, p. 24) occur by the hundreds, and one of them, an angular block (Pl. 2, fig. 1) 10 by 7 by 3½ feet, lies on the very summit, at an altitude of between 3700

and 3800 feet. The highest of these stones have been transported horizontally not less than 1000 feet toward the west and in the process have been lifted through a minimum of 100 feet. Coleman attempted to explain such occurrences as the result of topographic reversal. He believed that the immediately adjacent areas of granite on the east, which now lie about 100 feet lower than the summit of "First Mountain," formerly stood so high that great blocks of granite were shed downslope upon the lower metamorphic area that now constitutes "First Mountain." According to this hypothesis, the present erratics are mere residual weathering-cores of these granite blocks. To the writers this hypothesis is incredible because it demands that a higher area of granite should have been eroded to its present altitude, lower than metamorphic rocks of "First Mountain," during a time when isolated erratic blocks of granite have withstood long continued weathering.

Nor are these erratic granites the products of stream transport during the development of the preglacial erosion surface inferred by Alcock, because (1) many of them are too large to have been transported by this means, and (2) granitic rock can hardly be expected to have survived in firm fresh condition throughout Pleistocene time.

Hence the opinion is reached that these granitic boulders are glacial erratics and that as such they demonstrate glaciation up to at least 3700 feet (more than 200 feet higher than the highest striae seen). Further, they demonstrate movement westward (outward) radially away from Tabletop Mountain.

On other parts of Tabletop there are also many boulders and small stones of granitic and certain metamorphic rocks. Most of these are residual. On the other hand some of them may have been transported by glacier ice to their present positions, but they are so like the assemblage of local rocks that constitute the felsenmeer in which they occur that derivation as erratics cannot be proved.

The scarcity of stones of proved glacial origin on Tabletop is characteristic of the whole Gaspe Peninsula, as has been recognized by all geologists who have worked in Gaspe. This is partly because the drift, however scanty, includes a very large proportion of local rock types and partly because the dense cover of timber makes observation difficult, and stones of any kind are not easily seen except along valley floors where they have been reworked by streams. Other factors involved are discussed in a later part of this paper.

Broad glaciated surfaces.—Tabletop is marked in places both by rock basins and by broad smoothed undulating bedrock surfaces truncating joint structures. Neither of these features occurs on the highest sum-

mits, where only felsenmeer is in evidence, but both occur at various places elsewhere on Tabletop. Examples of broad glaciated surfaces are found east of the head of the Lac aux Américains cirque, at 3500 feet, and west of the head of the cirquelike valley of Ruisseau Auclair, at about 3300 feet. The lakes in the former locality have rock basins (Pl. 2, fig. 2), as have those one mile to the north (Coleman, 1922, p. 36). Some of the lakes 1 to 2 miles northwest of Mt. Jacques Cartier are rimmed by vegetation and may not lie in individual rock basins.

These surfaces exhibit no glacial polish in detail, for they have been roughened by weathering, yet the writers believe they are good evidence of glaciation.

Close relation of mantle to underlying rock.—In summit areas where the relation of the frost-weathered mantle (felsenmeer) to the underlying rocks can be seen, the lithologic character of the mantle changes rather abruptly at the contact between any two kinds of bedrock. If the summits had been undergoing weathering and mass-wasting since preglacial time the resulting downslope movements of the mantle should have distributed debris far downslope across geologic contacts. As this does not appear to have occurred, it may be reasoned that these processes have been in operation only a short time, prior to which any previous accumulation had been removed. This relationship constitutes strong indirect evidence that the preglacial mantle was removed by glaciation.

Summary.—The direct evidence of glaciation on Tabletop consists of broad glaciated rock surfaces, including rock basins, as high as 3500 feet, glacial striae as high as 3500 feet, and erratic stones at more than 3700 feet. Immediate indirect evidence consists of a close relation of coarse, frost-weathered mantle to the immediately underlying bedrock. Only three summits reach or exceed 4000 feet (Mt. Jacques Cartier and the unnamed summits immediately north and southwest of it, respectively). From these no direct evidence of glaciation has been obtained.[3]

[3] In discussing Mt. Washington, New Hampshire, and Mt. Katahdin, Maine, both of which were submerged beneath the last ice sheet, Antevs (1932, p. 7, 8) stresses the absence of a recognizable upper limit of glaciation as evidence that these peaks were glaciated. The glacial features, distinct on lower slopes, become less and less distinct with increasing altitude. However, this fact, although suggestive, is hardly evidence of glaciation of the summits. In the mountainous highland of northeastern Washington, where, according to observations made by the senior author, all gradations occur from summits once wholly covered by the last glacier ice to extensive nunataks, the upper limit of former glaciation is hardly anywhere distinct but has to be inferred from the altitudes of the highest foreign stones on the mountainsides. The authors see no reason why an ice sheet in a mountainous region should necessarily leave a well-defined record of the highest position of its surface.

CASE FOR NONGLACIATION

Coleman's (1922) conclusions on the glaciation of the high Shickshocks were as follows:

> "There is no evidence that glaciers worked on the highest parts of the Shickshock mountains, for their surfaces consist of loose blocks of the underlying rock, seldom showing any displacement; and no striated surfaces or boulder clay have been found above 2,500 feet." (p. 14.)
>
> "It [Mount Albert] resembles some tablelands in the Torngat mountains of Labrador and seems never to have been crossed by ice, though ice may have moved outwards from it." (p. 31.)
>
> "The block-covered summits of the hills and mountains of Tabletop and the entire absence of boulder clay, moraines, or smoothed and striated surfaces argue strongly against an outward movement of ice from the centre; and these features, along with the sharp-edged, northwestward-facing cliffs at the edge of the table, provide conclusive evidence against the supposition that the Labrador ice-sheet ever crossed the Shickshocks." (p. 37.)

The writers do not agree with these statements, in part as to fact and in part as to interpretation. With regard to Mount Albert, the presence on its plateau surface of stones derived from lower altitudes proves that it was glaciated, though the possibility is not excluded that the glacier was a local ice cap centered over Tabletop. In the writers' opinion Tabletop itself bears direct evidence of glaciation up to an altitude of at least 3700 feet. Of the several points on Tabletop that stand higher than this, only two were examined, owing to unfavorable weather. Mt. Jacques Cartier, the highest point of all, was examined in great detail. Its broad domelike surface is covered with local rock fragments arranged in polygons at its summit and in stone-stripes and broad felsenmeer on its flanks, and despite intensive search no stones recognizably foreign to the granitic bedrocks of this dome were found upon it.

Referring back to the statements quoted from Coleman, we find that his opinion that the high Shickshocks were not glaciated was based on (1) lack of glacial features at altitudes of more than 2500 feet, and (2) the presence of felsenmeer composed of local rocks in the summit areas. It has been shown that glacial features do occur on the summit of Mount Albert, and on Tabletop up to an altitude of at least 3700 feet. This leaves in question only the 500-foot zone between the highest glacial features and the highest summit (Mt. Jacques Cartier). This zone is indeed largely covered by felsenmeer composed of local rocks (Pl. 2, fig. 3) but, as pointed out by Alcock (1935, p. 125) and believed by the present writers, this is not evidence against glaciation because similar felsenmeer areas occur in places where glaciation has been proved by direct evidence.

One of these is Mt. Washington (6288 feet), dominating the Presidential Range in New Hampshire. Mt. Washington has been held by C. H. Hitchcock (1904, p. 83), J. W. Goldthwait (1913), Antevs (1932, p.

1-7), and R. P. Goldthwait (1940, p. 17-19) on a basis of rare though unmistakable striae, till, and erratic stones, to have been overtopped by glacier ice within late Pleistocene time, and this despite the fact that extensive felsenmeer, stone-stripes, and stone-polygons have developed on its summit area since the mountain emerged from beneath the ice. Upon leaving the Shickshocks, the writers visited Mt. Washington in order to make direct comparisons. The similarity, both as to glacial features where they occur and as to felsenmeer, is striking. The features on Mt. Washington prove that extensive frost wedging and mass-wasting have occurred there since the last glaciation.

However, Mt. Washington is not the only place where extensive frost wedging has produced abundant felsenmeer within postglacial time. The top of Mt. Katahdin in Maine is extensively covered with felsenmeer, yet Tarr (1900, p. 436) found foreign stones on the highest summit (5267 feet) and thereby proved it had been submerged beneath an ice sheet. R. P. Goldthwait (unpublished communication) cites additional proof of the same character. Tarr recognized that, even were the foreign stones lacking, ruggedness of the summit would not prove nonglaciation. His argument was threefold: (1) The mountain projected high into the glacier, well above the zone of maximum glacial erosion. (2) The mountain was a nunatak for a time during deglaciation; this time must be added to the postglacial time available for frost-splitting action. (3) High altitude and position above timber line provide optimum conditions for intensive postglacial sharpening of the summit by frost.

Again, the high summits of the Long Range of Newfoundland bear abundant evidence of glaciation (MacClintock and Twenhofel, 1940, p. 1732-1739) despite the fact that Coleman (1926) considered the summits to record ancient glaciation or none at all. The Gros Morne (2651 feet), one of the highest summits in Newfoundland, was climbed in 1939 by the senior author of the present paper. Its broad domelike surface is completely mantled with a felsenmeer of sharp, ragged fragments of the local quartzite; yet scattered through the felsenmeer are scores of gneisses, dark igneous rocks, and other foreign stones and boulders, one of them, at the very summit, being 4 feet in diameter. The felsenmeer could have been developed only since glaciation.

The mountains of northern Labrador exhibit similar features. In the Torngat region, where the highest peaks slightly exceed 5000 feet, Bell (1884, p. 14DD) found that "the mountains around Nachvak are steep, rough sided, peaked and serrated, and have no appearance of having been glaciated, excepting close to sea-level." Daly (1902, p. 247-251) observed no drift higher than 2050 feet and suspected that no glaciation had occurred at more than 2100 feet. Coleman (1921, p. 26), on

similar evidence, went further and stated that the northeastern peninsula of Labrador had not been covered by an ice sheet. More recently Odell (1938, p. 205) found ice-abraded bedrock high on several mountains in the Torngat Range, the highest (perhaps not latest Wisconsin in date) being at an altitude of 4700 feet. Odell does not state whether these occurrences can be explained by radial flow of former local glacier ice or whether they demand the Labradorian ice sheet itself. In Odell's opinion the presence of frost-sharpened peaks and an extensive felsenmeer is not inconsistent with the concept of former ice-sheet glaciation because such features are present in parts of East Greenland and Spitsbergen that were certainly glaciated in late glacial or even later time. Similar conditions are known in Norway (Werenskiold, 1915, p. 359-360; Fig. 2).

These facts and opinions make it clear that postglacial frost splitting and mass-wasting in favorable localities have led to the development of extensive felsenmeer. The writers know of no reason why, analogously, this process should not have operated on the Shickshock summits during the same bracket of time. So far as the writers can learn, no argument supporting a contrary view has appeared in the literature. Until such argument has been advanced, the presence of felsenmeer on the Shickshock summits cannot be accepted as evidence that the area between 3700 feet and 4230 feet—the highest point—has not been glaciated.

DATE OF GLACIATION

The time elapsed since the glaciation of any district is estimated ordinarily by the amount of chemical decomposition of the drift or of glaciated bedrock, and by the estimated probable rate at which decomposition takes place.

The basis for estimating the date of glaciation of the high Shickshocks is not very satisfactory. It may be stated that:

(1) Striated surfaces on lime-silicate rocks are not notably dimmed by weathering processes.

(2) Mechanical weathering (chiefly frost wedging) is strongly in evidence on the summits.

(3) Recognizable chemical decomposition is confined to faint limonitization of some angular blocks of dark rocks to a maximum depth of about $\frac{1}{8}$ inch.

(4) Some boulders of coarse-grained granite have lost their corners and have been reduced to a subrounded shape.

(5) The centers of stone-polygons on the summit of Mt. Jacques Cartier consist of clayey silt and stones to a depth in one place of at

least 5 feet. This material is thought to be a product of decomposition of the underlying granite.

(6) Frost wedging is believed to be rapid at this latitude and altitude. The frequency with which the temperature fluctuates through the freezing point is probably great.

(7) Fog and mist are prevalent on the higher parts of Mount Albert and Tabletop during the summer. (6) and (7) are believed to be an adequate explanation of the observed disintegration and decomposition of the granite.

(8) Many rock-basin lakes are not yet filled with peat or with clastic sediment.

In view of these statements the writers believe that the glaciation of the high Shickshocks could have occurred in Wisconsin time and that no evidence is at hand to support the belief that the glaciation dates from pre-Wisconsin time.

SOURCE OF GLACIER

Coleman, Alcock, Jones, and the present writers agree that local glaciers radiated from Tabletop and Mount Albert in late glacial time. This is shown by low-level cirques in the outer slopes of these two plateaulike masses and by evidence of the radial transport of drift from them. The altitude of the upper surface of this radially flowing ice cannot be inferred from the evidence collected. At one time the ice may have been an extensive cap completely burying the high Shickshocks. At its maximum it must have been at least high enough to overwhelm "First Mountain" on the west side of Tabletop, the broad central area of the plateau thus being covered. Above such an ice cap Mt. Jacques Cartier could have stood, if at all, no more than 400 to 500 feet. At a later time it may have thinned and become segmented into several independent glaciers. That this radial flow postdated any invasion by ice from north of the St. Lawrence is suggested by a morainelike feature at Lac aux Américains, whose form, convex downvalley, points to its construction by a glacier flowing westward from Tabletop.

That the high Shickshocks were overtopped by the Labradorian ice sheet cannot be positively affirmed from direct evidence, which must consist of the presence of stones derived from the region of northern Quebec, resting on the highest parts of the Shickshocks. However, the following lines of indirect evidence have been advanced for believing that the Labradorian ice sheet did actually overwhelm the Shickshocks:

(1) According to Alcock (1935, p. 125),

"The whole plateau country with elevations up to 1,800 feet shows drift, and the Shickshock region appears to be too small to have supplied the drift that spread

out so far. The evidence of Labradorian erratics [*cf.* Jones and McGerrigle, 1939, p. 31] also appears to demand that the ice sheet crossed the peninsula. It either did this or lapped around the Shickshocks."

(2) Also according to Alcock (1935, p. 125),

"A fact that also must always be considered in connexion with the subject of glaciation in the Shickshocks is that mountains considerably higher than any in that range and much farther away from the center of ice dispersal show evidence of having been crossed by the continental glacier. Mount Katahdin in Maine, elevation 5,150 feet, and Mount Washington in the Presidential range of the White Mountains of New Hampshire, elevation 6,293 feet, both show glaciated surfaces and numerous foreign erratics consisting of granites, gneisses, etc. With a continental glacier crossing these peaks it would seem that a thickness of ice of some 2,000 or 3,000 feet must at that time have covered the highest Shickshocks."

At first sight this argument seems convincing. However, the Shickshocks do not lie in a line connecting the White Mountains or Mt. Katahdin areas with the supposed centers of Labradorian ice dispersal; rather, the Shickshocks are on a radius removed about 20° from a radius through Mt. Katahdin. Thus thick ice could have overwhelmed Katahdin when relatively thin ice existed along the radius toward the Shickshocks. Such a condition might have been brought about because the St. Lawrence lowland, adjacent to the Shickshocks, would have allowed easier and more rapid discharge of ice toward the eastern margin of the ice sheet than toward the south. The present writers do not argue that this must have been the case. The suggestion is made merely to show that Labradorian glaciation of Mt. Katahdin and the White Mountains does not necessarily prove that the Shickshocks were ever overwhelmed by Labradorian ice.

(3) The fact that erratics of possible Labradorian origin occur on the summit plateau of Mount Albert might be taken as evidence that Labradorian ice had crossed this area. However, as already pointed out this evidence is equivocal.

As indicated, these three lines of indirect evidence do not demonstrate that Labradorian ice crossed the Shickshocks, although they are consistent with the possibility that this occurred.

The chief reason why indirect evidence has to be relied on is the rapid upward diminution of drift in the high Shickshocks. This is not peculiar to the Shickshocks but is characteristic of many glaciated regions of great relief and steep slopes. Explanation probably lies in four factors, two of which are inherent in the original distribution of the drift, whereas the other two are inherent in secondary redistribution:

A. Original deposition.

(1) There is an upward component of basal flow in ice that encounters a steep adverse slope. This retards basal flow and thus tends to cause

deposition near the base of the slope, leaving little drift to be deposited at higher positions.

(2) The greater velocity of ice flow in valleys as compared with adjacent intervalley areas tends to concentrate drift-laden ice in the valleys. As a result, when deglaciation occurs there is greater deposition of drift in the valleys than on the intervalley areas.

B. Secondary redistribution.

(3) It is clear, as pointed out in the present paper, that in late glacial time the high Shickshocks constituted a center of radial glacial outflow. This movement must have carried outward and downward a large part of whatever drift had been left at high levels by any pre-existing larger ice sheet. This factor could operate only in those highland areas that were subject to local nourishment and outflow of glacier ice during or following general regional deglaciation.

(4) Mass-wasting and stream erosion following deglaciation has been a powerful factor in many places, tending to emphasize the originally very unequal vertical distribution of the drift. On many steep slopes mass-wasting processes are believed to be competent to remove all of the drift deposited on them during glaciation.

Radial outflow of ice from Tabletop has already been shown to have occurred in late glacial time. That this radial outflow was not confined to the cirque-headed valleys that cut the outer slopes of the Tabletop plateau but occurred on the plateau summit up to an altitude of more than 3700 feet is shown by the granite erratics derived from the central rock mass of Tabletop but now resting on the metamorphic rocks of the western margin. The extent and thickness of the Labradorian ice sheet when this radial flow occurred is not known, except that it must have been thick enough to allow a local gradient, sufficiently steep to enable ice to flow radially away from Tabletop.

REFERENCES CITED

Alcock, F. J. (1926) *Mount Albert map area, Quebec,* Geol. Survey Canada, Mem. 144.

——— (1928) *Zinc-lead field of central Gaspé, Quebec,* Geol. Survey Canada, Summ. Rept., 1927, Pt. C, p. 27-46.

——— (1935) *Geology of Chaleur Bay Region,* Geol. Survey Canada, Mem. 183.

Antevs, Ernst (1932) *Alpine zone of Mt. Washington range,* Merrill & Webber Co., Auburn, Maine.

Bell, Robert (1885) *Observations on the geology, mineralogy, zoology and botany of the Labrador coast, Hudson's Strait and Bay,* Geol. Survey Canada, Rept. Prog., 1882-1883-1884, p. 1-62DD.

Coleman, A. P. (1921) *Northeastern part of Labrador, and New Quebec,* Geol. Survey Canada, Mem. 124.

Coleman, A. P. (1922) *Physiography and glacial geology of Gaspé peninsula, Quebec,* Geol. Survey Canada, Geol. Ser. no. 41.

——— (1926) *The Pleistocene of Newfoundland,* Jour. Geol., vol. 34, p. 193-223.

Daly, R. A. (1902) *The geology of the northeast coast of Labrador,* Bull. Mus. Comp. Zoöl., Harvard Coll., vol. 38, p. 205-270.

Fernald, M. L. (1925) *Persistence of plants in unglaciated areas of boreal America,* Am. Acad. Arts Sci., Mem., vol. 15, no. 3, p. 237-342 (quarto). (Also Mem. Gray Herb., Harvard Univ., no. 2.)

Goldthwait, J. W. (1913) *Following the trail of ice sheet and valley glacier on the Presidential Range,* Appalachia, vol. 13, p. 1-23.

Goldthwait, R. P. (1940) *Geology of the Presidential Range,* N. H. Acad. Sci., Bull., no. 1.

Hitchcock, C. H. (1904) *Glaciation of the Green Mountain Range,* Rept. Vt. St. Geol. 1903-1904, p. 67-85.

Jones, I. W. (1933) *The Tabletop map-area, Gaspé Peninsula,* Quebec Bur. Mines, Ann. Rept. 1932, Pt. D, p. 3-32.

——— and McGerrigle, H. W. (1939) *Report of geology on parts of eastern Gaspe,* Quebec Bur. Mines, P. R. no. 130 (quarto).

MacClintock, Paul, and Twenhofel, W. H. (1940) *Wisconsin glaciation of Newfoundland,* Geol. Soc. Am., Bull., vol. 51, p. 1729-1756.

Mailhiot, Adhémar (1919) *Geology of Mount Albert, County of Gaspé, P. Q.,* Quebec Bur. Mines (Dept. Colonization, Mines, Fisheries), Rept. Mining Oper. (1918) p. 146-151.

Odell, N. E. (1938) *The Great Ice Age and its effects,* in Forbes, Alexander: *Northernmost Labrador mapped from the air,* Am. Geog. Soc., Spec. Publ., no. 22, p. 204-215.

Tarr, R. S. (1900) *Glaciation of Mount Katahdin, Maine,* Geol. Soc. Am. Bull., vol. 11, p. 433-448.

Werenskiold, W. (1915) *The surface of central Norway,* in Am. Geog. Soc., Memorial vol. Transc. Excur. 1912, New York, p. 357-365.

YALE UNIVERSITY, NEW HAVEN, CONN.; WESLEYAN UNIVERSITY, MIDDLETOWN, CONN.; YALE UNIVERSITY, NEW HAVEN, CONN.

MANUSCRIPT RECEIVED BY THE SECRETARY OF THE SOCIETY, MARCH 3, 1942.
PRESENTED BEFORE THE SOCIETY, DECEMBER 30, 1941.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 53, PP. 1231-1254, 2 FIGS. AUGUST 1, 1942

MISSISSIPPI DELTA—A STUDY IN ISOSTASY

BY ANDREW C. LAWSON

CONTENTS

	Page
Abstract	1231
Introduction	1232
Structure	1233
Surface slopes	1235
The isostatic problem	1235
Previous views on subsidence	1237
The initial load of sea water	1238
The supporting strength of the crust	1239
Seaward extension of the delta	1241
The bottom of the delta	1242
Stratification of the embankment	1244
Snout of the embankment	1245
Sloping sea floor	1246
Dimensions of the delta	1246
Supply of delta sediments	1247
Initial mean altitude of Mississippi basin	1248
Fluctuation of sea level	1248
End of the deltaic cycle	1251
Supplement	1253
References cited	1254

ILLUSTRATIONS

Figure		Page
1.	Diagram of structure of delta	1235
2.	Diagrammatic north-south cross section of the Mississippi delta	1247

ABSTRACT

The work of Barton, Russell, and the Geological Survey of Louisiana has shown that the Mississippi delta is over 30,000 feet thick and that it has subsided *pari passu* with the deposit of sediment. They have considered isostasy as an explanation of this subsidence, and Barton has rejected the explanation. Russell and others have not satisfied themselves that the subsidence and the great thickness of the delta are due to the maintenance of an isostatic balance which was disturbed by sedimentary loading. The reason for the failure to establish the operation of the principle of isostasy is due to the fact that the load of water displaced by the delta has been ignored as a primary factor in the problem. No cognizance has been taken of the depth of the sea into which the delta was built. In the present paper it is shown that a delta like that of the Mississippi built out into deep water will subside under sedimentary load and will maintain isostatic balance throughout its growth, up to a limit of thickness which is determined by the initial depth of water. Russell's view that the alternation of degradation and aggradation of the surface of the delta is due to the fall and rise of sea level caused by glaciation and deglaciation in Pleistocene time is accepted as proven.

INTRODUCTION

Geological processes characteristically recur in cycles. Every cycle has a beginning and an end, but the process, of which it is a manifestation, never ends if we consider the earth as a whole. The erosional cycle is typical of geological cycles in general. Continental erosion has been active since the beginning of recorded time and is still going on at a vigorous rate. In any cycle of this prolonged process its early stages pertain to the physiographically youthful aspect of the continental surface, when the mountains are new and the relief is high. Then erosion is at its maximum vigor. When the divides become low and the relief, within the hydrographic basin of a master stream, is reduced to so subdued a surface that the running water ceases to function as an agency of erosion, then the cycle is in its closing stage. It can be renewed only by a rejuvenation of the relief due to a recurrence of deformation of the earth's crust or to a lowering of sea level relatively to the continental surface.

Corresponding to the erosional cycle there is a depositional cycle in which, for every drainage system, the waste products of degradation are carried to the sea and there dumped in the form of a great embankment known as a delta. But the two cycles do not necessarily begin or end at the same time, at least insofar as we recognize in the delta the result and measure of deposition. Long before erosion in any hydrographic basin has completed its task and the surface of the land has been reduced to a peneplain, the delta may have reached the limits of growth. Those limits may be set by the strength of the earth's crust, if the sea in which the delta is initiated be deep enough. A delta is a load imposed locally upon the surface of the earth, which as it becomes *large* is accommodated by depression. The sinking is slow, and the upbuilding of the delta at its surface may, and usually does, keep pace with the lowering of its bottom. Thus the embankment, as it advances seaward, takes on the configuration of a slab, or formation, of sediment, resting on the immediately antecedent sea floor. The growth of the delta, or the increase of its dimensions, is thus downward as well as upward and seaward. The vertical dimension at any stage of its growth is the direct measure of the load imposed by the sedimentary process. From this must be deducted the load of sea water displaced. The top of the delta may be partly a land surface and partly subaqueous, and during growth this top is much more stable as to position than is the bottom. The latter slowly subsides under load stress, while the former, particularly as to its subaerial portion, is held relatively fixed by the sea level.

To accommodate the sinking of the embankment into the earth's crust there must be a flowage of rock in depth, out from beneath the depression,

equal in mass, per unit of area, to the load imposed. This flowage is mostly landward, whence the load is derived, and tends to balance the diminishing weight of the crust in that direction. When the depression under the delta becomes broad and the embankment of weak incoherent sediments, to a thickness of several miles, becomes an integral part of the crust, the latter enters upon a critical stage of the whole geologic process. The tangentially compressive stress hitherto dormant in the outer part of the earth becomes active, and the weakened crust, together with the contents of the depression, collapses in the form of a geosyncline. The strata of the great embankment are folded, and a mountain range is projected slowly and fitfully far above sea level. The depositional cycle has closed, and another has been inaugurated by the diversion of the river mouth to some other part of the continental coast.

If, however, the load stress, measured in terms of thickness of the delta, together with the corresponding thinning of the strong crust, be insufficient to induce this horizontal collapse, the embankment may grow seaward indefinitely and so greatly prolong the duration of the depositional cycle. As will appear in the sequel the limiting thickness of the delta is determined by the depth of the sea at the inauguration of delta construction. Thus only deltas built out into relatively deep seas develop the *loci* of orogenic axes.

It is the purpose of this paper to review somewhat more in detail the history of such a depositional cycle, making particular reference to the delta of the Mississippi River.

The delta is considered to have been controlled in its growth and to have had its thickness determined by the operation of the principle of isostasy. The effort of the paper is not so much to give a detailed account of the delta as to use it, notwithstanding some defects of data concerning it, as an illustration of how the isostatic mechanism works. In the discussion it is necessary either to make certain assumptions or to treat the problem algebraically. I have preferred to make the assumptions in order to present a concrete picture of this great geological process, feeling confident that other values, which might reasonably have been assumed, would not have changed the picture essentially, but would have resulted in merely numerical differences. The discussion throughout is deductive, and the fundamental assumption is that of the validity of the doctrine of isostasy.

STRUCTURE

Our knowledge of the structure of the great deltas now functioning in various parts of the world has come from a study of small embankments at the mouths of rivers discharging into lake basins. On the shores of

Lake Superior one may stand at the mouths of small streams and watch the process of delta building. The drainage of the hinterland of the lake laden with sand and gravel issues with great velocity from small gorges. The moment that the stream emerges from the confinement of the gorge, the water spreads, becomes shallow, loses velocity, and drops parts of its load. From this point forward to the lake shore the stream has a much lower gradient, across the surface of the embankment which it has built up. Not only is the gradient and therefore the velocity diminished, but the stream divides into several distributaries, which flow radially from near the mouth of the gorge to the curving front of the delta at the shore. These distributaries drop their load of detritus progressively over the top of the embankment as they flow to the lake. They also bifurcate and otherwise subdivide, and the position of their channels is very inconstant. They thus spread the stream detritus fairly evenly over the top surface of the delta. The only unevenness worthy of note is due to the fact that the maximum dropping of stream load is at the borders of the channels, so that the latter become situated on low ridges. It is this precarious perch of the channels which induces their frequent change of course. At the brink of the embankment we may observe its outer slope down to and into the water and see that it is much steeper than the depositional slope of the top of the embankment. But this steeper slope lakeward is also a depositional slope, and we may safely infer that it has been so since the early stages of delta building. That the deposits on the flatter top slope rest on the edges of the more steeply inclined beds is also a safe inference. Looking lakeward from the outer edge of the embankment it is easy to infer that, out beyond the toe of the inclined beds, deposits of fine material are being laid down in layers parallel with the lake bottom. As the embankment grows lakeward the steeply inclined beds encroach on these flat-lying deposits and bury them.

Thus, from what I have said of the mode of deposit in a functional Lake Superior delta, anyone might have inferred the structure of deltas in general. But this is not the way we learned how deltas were built. It required a *post-mortem* examination to give us that knowledge. G. K. Gilbert about 60 years ago studied the basin of Lake Bonneville and described for us (1882) the shore features of the lake. The waters of Lake Bonneville at one time stood about 1000 feet above the level of the present Great Salt Lake. As the lake basin was emptied, down to the existing residual lake, the shore features of several of its stages were left high and dry on the mountain slopes of the desert. Among these were many deltas, and, as the surface of Lake Bonneville receded below the shore lines of which they formed a notable part, the streams cut trenches through them from top to bottom. This exposure of the whole delta

structure enabled Gilbert to analyze and describe it in detail. He observed the threefold make-up of the embankments to which I have alluded in the Lake Superior deltas and he gave names to their constituent parts. He called the deposits which form the nearly flat surface the *top-set* beds, the more steeply inclined beds beneath these the *fore-set* beds, and the nearly flat beds below these the *bottom-set* beds. He also showed the

FIGURE 1.—*Diagram of structure of delta*

relationship of these three structural elements in a diagram similar to that which is shown in Figure 1.

SURFACE SLOPES

This brief account of the small delta, based on Gilbert's classic work on the shore features of Lake Bonneville, serves equally well for the structure of the great deltas of the earth. The vast alluvial plain of the delta of the Mississippi, 386 km. in width, is continued seaward as a subaqueous slope for 129 km. to the 100-fathom contour, and the two together represent the expanse of the top-set beds. The subaerial portion of this expanse is nearly flat, and the subaqueous portion has a gradient of 1 in 710. Below the 100-fathom contour the outer slope of the delta has a gradient of 1 in 69 for another 195 km., down to the —3 km. (1640 fathom) contour. This is the region of deposition of the fore-set beds. For the next 180 km. seaward, down to contour —3.65 km. (2000 fathoms), the bottom-set beds have a gradient of 1 in 225. There can be little doubt but that, throughout the growth of the delta, the top-set beds have always been deposited across the upper edges of the fore-set beds at the front of the delta and that the latter have always extended out over the bottom-set beds.

THE ISOSTATIC PROBLEM

The most important feature of the Mississippi delta, from a geological point of view, is the great thickness of the embankment. It is recognized by geologists who know the delta best that the magnitude of its vertical dimension is due to the sinking of the old sea floor upon which the embankment rests. This subsidence has been a slow process and is probably still going on in the region of active growth of the embankment. It appears, moreover, to have been concomitant with the progress of growth without serious interruption from late Cretaceous time to the present. It is a short and easy step from the recognition of this association of

sedimentation and subsidence to suppose that the sinking of the sea floor is due to the load of sediment which has been brought to it and dumped upon it by the great river—in other words, that the subsidence is isostatic. But isostasy means more than the depression of the earth's crust by a load imposed upon it. Under such load, whether it be due to continental glaciers, thrusting, or the growth of embankments, the sinking is an expression of disturbance of balance. To compensate for the newly imposed load heavy rock in the depth of the earth must flow away to other regions and so make depression possible; and the outflow not only creates depression, it maintains balance, or constancy of gravitative stress, in the column of the earth's crust beneath the disturbing load. For the region on which the delta is being built we may assume, for convenience in discussion, that at the beginning of the deltaic cycle it was in stable isostatic balance.

When we attempt to apply the isostatic mechanism to a sinking delta embankment, however, we encounter difficulties which inhere in the diversity between the density of the load imposed and that of the rock which flows away in depths. The latter has a specific gravity of about 3.3, while the former, even after compaction, does not exceed about 2.4. Thus a layer added to the delta, having a thickness of 3.3, will be balanced by the removal of a layer of heavy rock 2.4 thick. The outflowage of the heavy rock creates a depression which permits the bottom of the delta to sink 2.4, but its top rises by deposition $3.3 - 2.4 = .9$. It is obvious that the top of the delta is limited to the proximity of sea level and cannot rise higher by river deposit. The incompetence of the river to place its load above the level of its flood plain, at or close to sea level, appears to forever inhibit the embankment from depressing its floor beyond a very small initial amount.

But if we suppose a considerable depth of water to have been displaced by the delta, then the mechanism works. In general, if x be the thickness of the embankment in kilometers, when it comes to final balance, and d the initial depth of water, the sea level being supposed constant throughout the deltaic cycle,

$$\text{then } 2.4\,x - 1.026\,d = 3.3\,(x - d)$$
$$\text{whence } x = 2.52\dot{6}\,d$$

Any addition to the thickness in excess of $2.52\dot{6}\,d$ kilometers would cause an isostatic depression less than the added thickness so that the latter would pile up above sea level. The river cannot so pile its deposit, and $2.52\dot{6}\,d$ kilometers is the limiting value for the thickness of the delta.

The assumption of a constant sea level is, however, inadmissible. We must consider the effect of the change of sea level on the growth of the delta. The quantity of sediment derived from all continental erosion and

delivered to the sea during the time of growth of the Mississippi delta is so great that it has raised the level of the ocean by an amount that cannot be ignored. Elsewhere (Lawson, 1938) I have estimated that the rise of sea level due to this cause, within the time mentioned, is .6 km. This additional height of sea level means that the surface of the ocean is so much farther from the center of the earth than it was at the beginning of the deltaic cycle. During the rise the continental surfaces have suffered degradation. For every layer having a mean thickness of 3.3 removed a corresponding thickness of 2.67 has been added in depth, so that the distance from the continental surfaces to the center of the earth has been diminishing, while the sea surface has been rising. Should the two surfaces, thus approaching each other in position, ever coincide—*i.e.*, should extensive continental surfaces be reduced to peneplains—the rise of sea level would thereby be checked. The increase of distance of the sea surface from the center of the earth has come about slowly with the growth of the delta embankment and does not represent displaced sea water as does the initial depth of sea displaced in that growth. It does, however, determine an upward growth of the delta, also of .6 km. or 1968 feet. This greatly modifies the estimate just made for the limiting thickness of the delta.

If, as before, x be the thickness of the embankment at final isostatic balance, d the initial depth of the sea in which it was built, and k the rise of sea level in the deltaic cycle to the present time, then

$$2.4\,x - 1.026\,d = 3.3\,(x\text{-}d\text{-}k)$$
$$\text{whence } x = 2.52\dot{6}\,d + 3.6\,k$$

In the case under consideration $k = .6$ km. and $x = 2.52\dot{6}\,d + 2.2$.

PREVIOUS VIEWS ON SUBSIDENCE

In two excellent papers (Barton, *et al.*, 1933; Russell, 1940) dealing with the subsidence of the Mississippi delta this question of the initial depth of water is not discussed, although it is apparent that the authors consider the water to have been so shallow that they are unable to explain the great subsidence, which they make known, as a manifestation of the operation of isostasy. Barton and his associates state that since the beginning of the Tertiary there has been continuing progressive subsidence. Also they say (p. 1458):

"But isostatically, the Gulf Coast geosyncline must be, and for a long time must have been, negatively out of equilibrium. Subsidence continued, however, and presumably must have increased the lack of isostatic equilibrium, as the progressive depression of the basement has increased the negative gravity anomaly. The movement, therefore, has been the reverse of what would be expected from the theory of isostasy. . . . The subsidence, therefore, seems more probably to be the effect

of the sedimentation and to have tended to compensate it. But the subsidence cannot be the effect of a movement toward isostatic equilibrium under the effect of extra load of sediments. Superficial yielding of the basement under the weight of the extra load of the sediments seemingly must have taken place without regard to the isostatic relations."

Throughout the discussion from which these quotations are taken there is no reference to the depth of water displaced by the advancing embankment. The implication is that the authors considered it to be negligible, but they do not say even that much, although the water displaced is an important factor in the operation of the isostatic mechanism.

In Russell's (1940, p. 1227, 1228) fine paper, he says:

"It seems difficult, if not impossible, to accept Louisiana evidence without coming to the final conclusion that the Gulf Coast Geosyncline is the result of sedimentary loading. . . . The competence of sedimentary loading to depress material beneath and thus permit additional load at the original site seems clearly established by the lithologic and structural record."

But nowhere does he raise the question of the depth of water in which the delta embankment was built out until it attained mass enough to initiate depression. He seems, however, to be impressed with the fact that: "Continental, near shore, and shallow water sediments are typical of its whole Cenozoic column." Thus, although Russell recognizes the isostatic implications of depression under load, he, too, seems to regard the initial depth of water as negligible.

THE INITIAL LOAD OF SEA WATER

Like Barton and Russell I am impressed by the suggestion that, in the sinking of the Mississippi delta under its own load, we have an example of the operation of the principle of isostasy. I am also aware, as they seem to be without saying so, that, if the delta started in shallow water, we cannot invoke isostasy to explain the sinking. But, as far as I know, there is no reason whatever for the supposition that the delta was built out into shallow water. The counter assumption is that, when the delta started, there was a heavy load of sea water on the sea floor which was gradually displaced as the delta embankment extended out from the original shore. For the known thickness of the delta, 30,000+ feet south of New Orleans, the depth of such a sea could not have been less than 2.75 km. to provide the necessary mechanism for isostatic subsidence, which seems to exist in that part of the delta. Since, from the formula given above, if x be the required depth,

$$2.52\dot{6}\,x + 2.2 = 9.146 \text{ km. (30,000 feet)}$$
$$\text{whence } x = 2.75 \text{ km.}$$

But the 30,000 feet for the thickness of the delta to the south of New Orleans is a minimum value. I propose, therefore, to assume that the

initial depth of sea in the northern part of the Gulf of Mexico was 4 km. and that the region was in isostatic equilibrium when that was the case. This estimate accords with the fact that soundings off coast lines aggregating 30,000 km. in Africa, India, Australia, North America, and South America show that in 61 profiles the —4 km. contour is distant from the shore on an average only 52 km.

There is nothing in the assumption which precludes the existence or, indeed, the prevalence, of shallow-water deposits throughout the entire geological column represented in the delta. Shallow-water sediments are a prime characteristic of all deltas, whether they are built out into deep water or shallow water; and nothing can be inferred from their presence in a depressed delta as to the depth of water antecedent to its displacement by the embankment.

For an assumed initial depth of water of 4 km., on a level sea floor, and a rise of sea level of .6 km. during the deltaic cycle, the limiting thickness of the delta, for all that extensive portion of it which is at, or slightly above, sea level, is $2.52\dot{6} \times 4 + 2.2 = 12.3$ km., provided, as already stipulated, that the earth's crust has in itself no supporting strength. But probably in no case would the thickness reach that limit. The load has to accumulate to a certain critical thickness, determined in large measure by the area over which it is spread, in order to inaugurate subsidence. This means that the more rigid crust, above the asthenosphere, has a certain very limited degree of supporting strength, inherent either in the arch of the earth's spherical structure, or in the frictional resistance to flow in the asthenosphere, when a layer of heavy rock is forced to evacuate its position, and so make subsidence possible. But, as the process of evacuation is excessively slow, the frictional resistance should be small. But, however small, it is probably the more important factor in giving to the crust its supporting strength in the early stages of loading.

THE SUPPORTING STRENGTH OF THE CRUST

So far as I am aware, we have no positive measure of the supporting strength of the earth's crust under load, apart from isostatic balance. The nearest approach to such a measure is Gilbert's (1890) suggestion that the emptying of Lake Bonneville caused an arching of the region by removal of load. In this case, accepting the suggestion, the Bonneville shore line, at an altitude of about .3 km. above the present Great Salt Lake, was arched when the water level dropped to the level of the Provo stage of the Lake, and then to that of Great Salt Lake. At the Provo stage, by reason of a relatively constant outlet, the water stood for a long time and developed magnificent shore features. But when the water

in the basin evaporated down to the level of Great Salt Lake the abandoned Provo shore, about .18 km. above the present Lake, was also arched. The arching of the Provo shore was, however, not so acute as that of the Bonneville shore by removal of the load of water. We may infer, therefore, that a load of .18 km. of water is sufficient to depress the earth's crust in that region. In the absence of other data we may consider this load, if applied as broadly as it was in the Bonneville basin, as a limiting measure of the supporting strength of the crust. That is to say, any load equal to or greater than a prism of water .18 km. deep will depress the crust.

But even a large delta is a local affair, and the depression which accommodates it has traceable boundaries against the more stable region surrounding the embankment. It is difficult to avoid the conclusion that these boundaries, particularly those between it and the continental area, are zones of distributed shear, within which the subsidence is to some extent resisted by the strength of the crust.

Since we have no very definite measure of the competence of the crust, apart from isostasy, to support load we may to advantage compare two hypothetical cases. In the first the crust under the delta is supposed to have no supporting strength, or, to put it another way, the whole of the sedimentary load is effective for depression. Then, at final balance, when the embankment ceases to sink, if x be the total amount of depression,

$$2.4\ (x + 4.6) - 4 \times 1.026 = 3.3x.$$
$$\text{whence } x = 7.7.$$

The thickness of the embankment is then

$$7.7 + 4.6 = 12.3 \text{ km.}$$

In the second case a thickness y is assumed to accumulate on the sea floor before depression sets in. If x is the depression in kilometers, at final balance,

$$2.4\ (x + 4.6 - y) - 4 \times 1.026 = 3.3x$$
$$\text{whence } x = 7.7 - 2.6y.$$

That is, for every 1 km. of load ineffective for depression the subsidence at final balance is 2.6 km. less than in the case where the whole sedimentary load is so effective.

If, now, we apply to the Gulf of Mexico the Bonneville measure of the supporting strength of the crust, and take the maximum value for the load necessary to depress the crust, the equivalent of .18 km. of water is .131 km. of delta sediment after compaction and correction for

the levitation of sea water displaced. Then the depression at final balance is $7.7 - 2.6 \times .131 = 7.35$ km.; and the thickness of the delta is $7.35 + 4.6 = 11.95$ km.—let us say, 12 km.

The only possibility of the realization of this hypothetical load, which may have been imposed on the sea floor without depressing it, is in the region of the slope of the fore-set beds, as it was in the early history of the delta. Today that slope is but little less than 200 km. wide, from north to south, and its upper or landward edge is over 500 km. south of the original shore line at which it started. In its very early stages this slope may have been somewhat steeper than at present, but it has always advanced seaward approximately parallel to itself, and, therefore, of course always risen by accretion parallel to itself. The sea floor immediately underneath this slope would, even in the very early stages of delta building, first feel the effect of loading in excess of support at a short distance seaward of the northern limits of the embankment. Subsidence once started would proceed steadily with accretion at the top in the form of fore-set followed by top-set beds; and the area of subsidence would widen with the seaward advance of the embankment. The limit of subsidence would have been reached when the sea floor, on which the delta rests, had been depressed 7.35 km. below the position at which it was when delta building started, reckoning positions with reference to the center of the earth. For at this depth the position of the bottom of the embankment would be stabilized by the impossibility of adding more load at the top.

SEAWARD EXTENSION OF THE DELTA

Thus the sinking of the site of the delta began at a very early stage of its growth, and, from that beginning to the final stage of stability, subsidence proceeded steadily with growth at the top. In this growth it is not clear to what extent the supporting strength of the basement, up to a limit of less than that represented by a load equivalent to .18 km. of water, continued to operate after that limit was overtaken by increasing load; but it probably represents a permanent stress difference in the isostatic mechanism. The stress difference being constant throughout subsidence, the latter should have been at a constant rate in any given vertical locus of sinking. Apart from this stress difference we may say that, at every stage of the subsidence, if we may speak of stages in a continuous process, the embankment has been in temporary isostatic balance. But it is pretty clear that the top of the delta could not permanently reach sea level in its upward growth until the bottom had reached a depth of 12 km. below sea level, at which depth it ceased to sink. This means that the shore line of the delta at any time is the

locus out to which it has attained permanent balance. Both Barton and Russell appear to be of the opinion that subsidence is still going on. If this be so, it means, under the supposition of an initial depth of water of 4 km., that the embankment has not yet attained its limiting thickness of 12 km. But, in view of what has just been said, it seems improbable that this is so for the subaerial portion of the top of the delta. It seems more probable, since the sinking began at a very early stage of its growth, that the process involved the seaward extension of an embankment of the greatest possible thickness at all stages of that extension, out to the shore of the delta for every one of those stages. If this view be correct, it follows that the evidence of sinking cited by Barton, Russell, and others in the middle part of the subaerial delta plain is not a manifestation of depression now going on in that region, but is ancient history. The only part of the embankment now sinking, or capable of sinking, is probably the outer or seaward part, having a thickness of less than 12 km. All that part north of the shore has probably attained the maximum possible thickness and is stabilized in position until there comes an appreciable rise in sea level, which will make additions of new top-set beds possible.

THE BOTTOM OF THE DELTA

Here we may briefly consider the configuration of the bottom of the delta embankment. The profile of the top of the embankment is definitely determined by its relation to sea level. The land portion of the top, particularly, is the flood plain of a great river. Even if, in the mechanism of isostatic adjustment of disturbed balance, there were any tendency to arch the surface the tendency would be ineffective against the independent counteraction of the river and its distributaries; and any tendency to sag would be promptly countered by the filling of the depression by the river. Sagging has been, however, a very notable characteristic of the delta. As the embankment extends seaward, under an upward limitation at sea level for its top, the load becomes greater than the crust can sustain, and it becomes thicker by the sag of the floor on which it rests. The detailed evidence for this is set forth by Barton and by Russell in the papers heretofore cited. That evidence is unequivocal and satisfactory in the highest degree as to the subsidence of the embankment progressively with its growth. The increment of thickness comes to the embankment at its top, but accommodation for that increment is provided by depression of its bottom. As to the profile of the bottom Russell speaks of the general structure as being ladle-shaped, and the implication of his description is that the bottom is convex downward. Barton's figures show the embankment lying in a rather acute V-shaped trough, although the acuteness is largely due to the exaggeration of the

vertical scale of the drawings. If the delta is in isostatic balance, and the initial depth of water was uniformly 4 km., then for the maximum possible thickness of the embankment—*viz.*, 12 km.—its bottom must be level. This condition could only have been arrived at slowly and progressively with the seaward growth of the embankment. The determining factor, at each and every stage of this outward growth, was isostatic balance at a thickness of 12 km. Beyond that thickness there was no tendency to further subsidence. The delta had then arrived at stability of position. This stabilization, as a dynamic process, would keep pace with the advance of the embankment seaward; and the bottom of the delta would be level out to a line parallel to and vertically below the present shore contour. From this line southward, the bottom of the delta would slope upward at a small angle. Beyond the line, southward, the embankment would be in unstable balance in the sense that it is still capable of taking further load, and so still further depressing the bottom upon which it rests. When the present fore-set slope, by continued accretion of sediments, becomes built up close to sea level and is buried by top-set beds at sea level, then the upward slope of the bottom of this portion of the delta will have sunk to a position 12 km. below sea level, and have become part of the level bottom of the embankment. For a rigid body this sinking would be effected by rotation on a pivotal line, at the southern extremity of the flat bottom of the older part of the delta, at a depth of 12 km. The angle of rotation would be that which the upsloping bottom of the now functioning fore-set portion of the embankment makes with the horizontal. This angle is 1°33′. But since the embankment is far from being rigid the sinking may possibly be effected by shear stress and corresponding strain, with small ruptures, distributed throughout the mass over which the load is becoming heavy enough to cause depression. Under the isostatic theory of deltaic subsidence this is about the only way in which a large embankment could advance seaward and maintain isostatic equilibrium throughout the process. Of course since a very early stage in the growth of the delta—*i.e.*, since the initiation of subsidence—the main embankment has had a seaward projection of less thickness, in the region of the fore-set and bottom-set beds, which gradually became incorporated in the main embankment as the latter advanced seaward. This incorporation was at all stages effected by addition of new sediments at the top and by the sinking of the whole enlarged mass by the rotation or distributed shearing above referred to. The final result, for the assumption of a uniformly deep sea to begin with, would be a level-topped, level-bottomed embankment in isostatic balance at the limiting thickness determined by the initial depth of water. At the front of this embankment, even in its final

stage, would project those thinner underloaded portions, where deposition was still active, in the form of subaqueous top-set, fore-set, and bottom-set beds, and where movements toward final balance by rotation or shearing were still going on.

STRATIFICATION OF THE EMBANKMENT

I have called the top of the delta from the original shore line out to the 100-fathom contour, a width of 515 km., the region of the top-set beds. The subaerial portion of this has according to Russell (1940, p. 1216) an air-line slope of .32 feet to the mile. The strata in part of the top-set region have, therefore, a depositional dip of .32 feet per mile, and in the subaqueous part, 129 km. wide, the depositional dip is 600 feet in 129 km., or 80.15 miles—that is, 7.4 feet per mile. But while, under the hypothesis of an initial depth of water of 4 km. and a rise of sea level of .6 km., since the beginning of the deltaic cycle, the embankment has a thickness of 12 km., at the present shore line, and is in final isostatic balance, incapable of further subsidence, it may sink further between the shore and the 100-fathom contour. This subsidence will range from nothing at the present shore line to .18 km. at the 100-fathom line, when the shore shall have extended out to where there is now 100 fathoms of water. As the subaerial beds, dipping seaward .32 feet per mile, extend out over the present subaqueous beds, with their dip of 7.4 feet per mile, there will be an apparent disconformity between the two sets of beds. When the lower set has been thus buried, out to the present position of the 100-fathom contour, the added load will have caused the embankment to subside to its final position of equilibrium with a thickness of 12 km. In this settlement the block of the embankment between the present shore and the present 100-fathom contour will have effectively rotated through an angle the tangent of which is $\frac{.18}{129}$. Due to this rotation of the block the buried beds near sea level which had a depositional dip of 7.4 feet to the mile will have had that dip increased to about 15 feet per mile when they come to final rest.

If, however, by reason of certain constant characteristics, the whole of the deposit in this interval be considered a single stratum, the latter will have thickened steadily by accretion at the top, as well as by subsidence. When the shore line had reached the locus of the present 100-fathom contour its thickness there would be .362 km. (*i.e.* depth of water .18 km. and amount of subsidence .182 km.). If today a thin layer of volcanic ash were deposited on the sea bottom between the present shore and the 100-fathom contour, it would become a marker for the bottom of a wedge-shaped stratum. The dip measured on this

marker would be .362 km. in 129 km., or 14.81 feet to the mile; and the top would of course be level.

On the fore-set slope, the depositional dip (gradient of slope) is $3 - .182 = 2.818$ km. in 195 km. or 76.3 feet per mile. But when, by seaward extension of the embankment, the 100-fathom contour moves out 195 km., and the fore-set beds are buried, a layer of ash, laid down now on the sea floor, would have a gradient or dip which is the sum of the present dip of the fore-set beds and their angular depression, that is, $76.3 + 135.6 = 211.9$ feet per mile. It is, however, improbable that these final relatively steep dips of the fore-set beds would ever be seen at the surface of the delta, although they might very well be encountered in bore holes penetrating the depth of the embankment. Except in cases of trenching, due to fall of sea level in glacial epochs, dips observable at the surface are those of the top-set beds, and the steeper of these would be marine strata, laid down off shore in shallow water and steepened by subsidence in the manner above indicated.

SNOUT OF THE EMBANKMENT

It has been pointed out that the embankment comes to balance when it is 12 km. thick, after allowing for a stress difference equivalent to a load of .131 km. of sediment. Following the seaward extension of the delta shore, the embankment progressively attains balance by continuous deposition of top-set beds in shallow water and corresponding depression. Beyond the shore the cross section of the embankment is determined by (1) the slope of the shallow-water bottom out to the brink of the fore-set slope, and the corresponding upward slope of its bottom to a locus immediately below that brink; (2) the relatively steep slope of the present sea floor where the fore-set beds are accumulating, and the corresponding upward slope of the old depressed sea floor below this; and (3) the flatter sloping sea floor in the region of bottom-set deposition, and the corresponding upward slope of the old depressed sea floor. These three elements of the profile of the embankment are all in temporary balance for the masses which they respectively contain. All three are, however, susceptible to further subsidence as depositional load is added to them, to the limit of 12 km. below present sea level, continuing the same horizontal bottom and uniformity of thickness as prevail to the landward of the existing delta shore. The configuration of the profile of the delta at all stages, after final balance has once been attained in its early growth, is that of a flat-bottomed frying pan, rather than a ladle. The profile is, however, asymmetric. The steep side is to the north, seaward of the original shore line. The less steeply inclined side of the pan is in three straight segments, nearly symmetrical in gradi-

ent to the subaqueous top-set, the fore-set, and the bottom-set slopes, respectively. In general the profile beyond the shore is angular both at top and bottom, tapering seaward. It may be called the snout of the delta. The length of the snout of the delta is greater than the main body of the delta now in permanent balance, but its volume is less. The delta grows by the gradual conversion of the snout into a prism uniformly 12 km. thick.

SLOPING SEA FLOOR

The discussion of the bottom of the delta has been limited thus far to conditions flowing from the supposition of uniform initial depth of water—*i.e.*, a level sea floor in the region now covered by the delta. But it is, of course, entirely possible that the sea floor may have had a down slope seaward. Let us suppose that a portion of the sea floor had a gradient of 1 per cent. At the upper end of this slope the initial depth of water may be taken at 4 km. Then at a distance of 100 km. seaward the depth would be 5 km. As the embankment advanced seaward the slope would be buried by sediment about as fast as the level sea bottom, but it would take longer for the top to arrive at sea level even if there were uniform subsidence. But when the thickness of embankment on the lower part of the slope became greater than on the more shallow part, as it would in the course of time, the increasing load would doubtless accelerate the rate of subsidence and retard the time for completion at the top. In this case, if the curve of acceleration were not rectilinear, increase of load might cause a curvature of the buried sea floor—*i.e.*, of the bottom of the embankment. But whether that bottom were curved or straight, the thickness of the out flowage of heavy rock in depth for compensation would be greater beneath the lower part of the slope than beneath the shallower part. Its upper boundary in profile at final balance would be symmetrical with the bottom of the embankment—*i.e.*, whereas the latter at final balance would slope down seaward, the former would slope upward.

DIMENSIONS OF THE DELTA

The boundaries of the delta cannot, of course, be stated with exactitude. In an attempt to estimate its approximate area the northern boundary, though rather sinuous, is taken to be at Lat. 32°30′. The southern limit is a fading boundary on the sea floor north of Sigsbee Deep, or perhaps in that depression, where the bottom is very flat. The 2000-fathom contour, on the north side of the deep, may be taken as a convenient and conservative limiting line. In Long. 91° W. this line coincides with Lat. 24°18′ N. The delta thus covers over 8° of latitude, its width being 879 km.

The land surface of the delta from its northern boundary to the shore, at about Lat. 29° N., measures 386 km. on a N.-S. line. From the shore to the 100-fathom contour, at Lat. 27°57′ N., the distance is 129 km. These two portions of the delta, totaling 515 km. in width, comprise the flat-lying top-set beds of the embankment. The slope from the 100-fathom contour to the 1640-fathom (—3 km.) contour, comprising

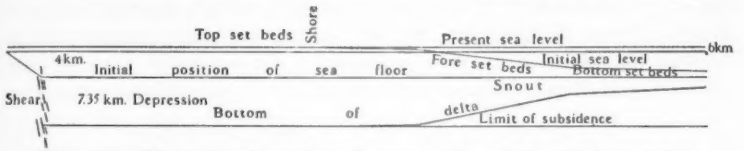


FIGURE 2.—*Diagrammatic north-south cross section of the Mississippi delta*

Showing initial and present positions of sea level, initial position of sea floor at an assumed depth of 4 kms., and the present bottom of the delta after depression to the limit of 7.35 kms. At the landward side of the depression a shear is suggested. Seaward of the shore, the delta, not yet depressed to the limit, tapers in the form of a snout.

the present surface of the fore-set beds, is 195 km. in width and has a gradient of 1 in 70. From the 1640-fathom to the 2000-fathom contour the width is 169 km., and the gradient is 1 in 256. This slope is in the region of the bottom-set beds, which fade insensibly into the deep-sea deposits.

Since the initial depth of water limits the thickness, and the depth is assumed to have been uniform over the northern part of the Gulf of Mexico, the bottom of the embankment is as level as the original sea floor. We are thus in a position to construct a hypothetical profile of the embankment, and such a profile is shown in Figure 2. The cross-sectional area of the profile is 8260 square kilometers. If this profile, which is drawn about on Long. 91° W., be taken as characteristic of the embankment throughout its length, we may arrive at approximate figure for the volume of the delta. The embankment extends from about Long. 87° W. to Long. 99° W., a distance of 830 statute miles or 1336 km. Russell speaks of "the geological processes which have been dominant since the late Mesozoic in forming a sedimentary section some 30,000 feet thick, extending coastwise for over a thousand miles." For the more limited figure which I have just given for the east-west extent of the embankment its volume is $8260 \times 1336 = 11{,}035{,}360$ cu. km.

SUPPLY OF DELTA SEDIMENTS

This large quantity of sediment has been brought to its present position by the agency chiefly of the Mississippi River. It is derived from the hydrographic basin of the river and represents about half the total erosional product. The work of Dole and Stabler (1909) supplement-

ent to the subaqueous top-set, the fore-set, and the bottom-set slopes, respectively. In general the profile beyond the shore is angular both at top and bottom, tapering seaward. It may be called the snout of the delta. The length of the snout of the delta is greater than the main body of the delta now in permanent balance, but its volume is less. The delta grows by the gradual conversion of the snout into a prism uniformly 12 km. thick.

SLOPING SEA FLOOR

The discussion of the bottom of the delta has been limited thus far to conditions flowing from the supposition of uniform initial depth of water—*i.e.*, a level sea floor in the region now covered by the delta. But it is, of course, entirely possible that the sea floor may have had a down slope seaward. Let us suppose that a portion of the sea floor had a gradient of 1 per cent. At the upper end of this slope the initial depth of water may be taken at 4 km. Then at a distance of 100 km. seaward the depth would be 5 km. As the embankment advanced seaward the slope would be buried by sediment about as fast as the level sea bottom, but it would take longer for the top to arrive at sea level even if there were uniform subsidence. But when the thickness of embankment on the lower part of the slope became greater than on the more shallow part, as it would in the course of time, the increasing load would doubtless accelerate the rate of subsidence and retard the time for completion at the top. In this case, if the curve of acceleration were not rectilinear, increase of load might cause a curvature of the buried sea floor—*i.e.*, of the bottom of the embankment. But whether that bottom were curved or straight, the thickness of the out flowage of heavy rock in depth for compensation would be greater beneath the lower part of the slope than beneath the shallower part. Its upper boundary in profile at final balance would be symmetrical with the bottom of the embankment—*i.e.*, whereas the latter at final balance would slope down seaward, the former would slope upward.

DIMENSIONS OF THE DELTA

The boundaries of the delta cannot, of course, be stated with exactitude. In an attempt to estimate its approximate area the northern boundary, though rather sinuous, is taken to be at Lat. 32°30′. The southern limit is a fading boundary on the sea floor north of Sigsbee Deep, or perhaps in that depression, where the bottom is very flat. The 2000-fathom contour, on the north side of the deep, may be taken as a convenient and conservative limiting line. In Long. 91° W. this line coincides with Lat. 24°18′ N. The delta thus covers over 8° of latitude, its width being 879 km.

The land surface of the delta from its northern boundary to the shore, at about Lat. 29° N., measures 386 km. on a N.-S. line. From the shore to the 100-fathom contour, at Lat. 27°57′ N., the distance is 129 km. These two portions of the delta, totaling 515 km. in width, comprise the flat-lying top-set beds of the embankment. The slope from the 100-fathom contour to the 1640-fathom (—3 km.) contour, comprising

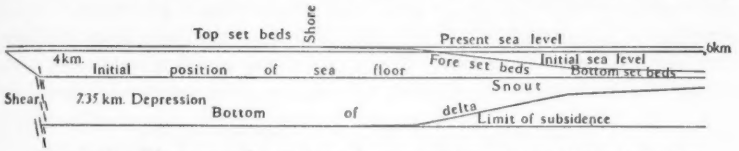


FIGURE 2.—*Diagrammatic north-south cross section of the Mississippi delta*

Showing initial and present positions of sea level, initial position of sea floor at an assumed depth of 4 kms., and the present bottom of the delta after depression to the limit of 7.35 kms. At the landward side of the depression a shear is suggested. Seaward of the shore, the delta, not yet depressed to the limit, tapers in the form of a snout.

the present surface of the fore-set beds, is 195 km. in width and has a gradient of 1 in 70. From the 1640-fathom to the 2000-fathom contour the width is 169 km., and the gradient is 1 in 256. This slope is in the region of the bottom-set beds, which fade insensibly into the deep-sea deposits.

Since the initial depth of water limits the thickness, and the depth is assumed to have been uniform over the northern part of the Gulf of Mexico, the bottom of the embankment is as level as the original sea floor. We are thus in a position to construct a hypothetical profile of the embankment, and such a profile is shown in Figure 2. The cross-sectional area of the profile is 8260 square kilometers. If this profile, which is drawn about on Long. 91° W., be taken as characteristic of the embankment throughout its length, we may arrive at approximate figure for the volume of the delta. The embankment extends from about Long. 87° W. to Long. 99° W., a distance of 830 statute miles or 1336 km. Russell speaks of "the geological processes which have been dominant since the late Mesozoic in forming a sedimentary section some 30,000 feet thick, extending coastwise for over a thousand miles." For the more limited figure which I have just given for the east-west extent of the embankment its volume is $8260 \times 1336 = 11{,}035{,}360$ cu. km.

SUPPLY OF DELTA SEDIMENTS

This large quantity of sediment has been brought to its present position by the agency chiefly of the Mississippi River. It is derived from the hydrographic basin of the river and represents about half the total erosional product. The work of Dole and Stabler (1909) supplement-

ing the earlier work of Humphreys and Abbott (1861) has shown that the surface of the basin of the Mississippi River is being denuded at the rate of .056 mm. per year.[1] Of this amount Dole and Stabler have shown that 25 per cent goes to the sea in solution and 75 per cent as solids. The present writer estimates that one-third of the solids goes to the ocean shores and bottom beyond the limits of the delta, and that two-thirds, or one-half of the total erosional product, goes to the construction of the delta.[2] Thus the total volume of the delta after correction for difference of density is somewhat greater than one half the volume removed from the area of the river basin. The erosional removal is

$$11{,}035{,}360 \times \frac{2.4}{2.67} \times 2 = 19{,}838{,}848 \text{ cu. km.}$$

Since the area of the basin is 3,276,274 sq. km., the thickness of the layer removed is $\frac{19{,}838{,}848}{3{,}276{,}274} = 6.03$ km., let us say 6 km.

INITIAL MEAN ALTITUDE OF MISSISSIPPI BASIN

For removal of a layer 6 km. thick from the area of the basin and the delivery of the same to the ocean, the isostatic rise of the eroded surface is $\frac{2.67}{3.3} \times 6 = 4.854$ km. The lowering of the surface of the basin is $6 - 4.854 = 1.146$ km. On the basis of a series of hypsometric sections drawn across the basin I have estimated that its mean altitude above present sea level is .80 km., or $.8 + .6 = 1.4$ km., above the position of sea level at the beginning of the deltaic cycle. From these data we may ascertain the mean altitude of the surface of the basin at the beginning of the cycle. Since the present mean altitude is .8 km. and the lowering of the surface is 1.146 km., the initial mean altitude above present sea level is $.8 + 1.146 = 1.946$ km., or $1.946 + .6 = 2.546$ km. above sea level as it was at the beginning of the deltaic cycle.

FLUCTUATION OF SEA LEVEL

If the region of the Mississippi delta were to be slowly uplifted 300 or 400 feet what would be the effect on its land surface? First of all the various rivers flowing to the sea across it, trunk stream and distributaries, would have their grades steepened and their velocity increased. As a consequence their corrasive power would be greatly enhanced. Every stream would tend to cut a trench for itself and, if the uplift were not too slow, would do so. No distributary could branch out from such a trench. The depth of the trench would be closely determined by the change in the relative position of sea surface and delta surface. The

[1] For summary of data, *see* Lawson (1934).

[2] There is, however, very little observational basis for this estimate.

antecedent surface of the delta would become a broad terrace traversed by the trenches. In time, when the streams had re-established their proper gradients in these trenches, they would proceed to widen them by lateral corrasion and evolve broad bottoms, thus accentuating the terraced aspect of the topography. Holding this picture in mind let us consider now what would happen if the region slowly subsided, till the former hypsometric relation of sea and land were re-established. How would the physiography of the delta surface change? The gradients of the streams in all these broad terraces would be lowered, first at their mouths by building little subdeltas, and thence gradually upstream. Eventually the trenches, no matter how broad, would be filled to the brim. Then every stream would build up levees a little higher than the plain on either side, and would sooner or later spill over to form new distributaries. The surface of the delta would have been restored to its normal character.

In the later geological history of the Mississippi delta, according to the brilliant work of R. J. Russell, this trenching of its surface to depths of from 100 to over 300 feet, followed by filling of the trenches, has occurred no less than five times. Frink (1941) confirms Russell's earlier statement regarding this five-fold recurrence of degradation and aggradation of the delta surface. Accepting Russell's reading of this deltaic history we might conclude from what I have said above that the region of the delta had been subjected to a five-fold recurrence of uplift and depression, thus recognizing an extraordinary mobility of the earth's crust. But our conclusion would be fallacious. The phenomena of the recurrence of alternate degradation and aggradation may equally well be explained by the rise and fall of sea level. Students of continental glaciation are well aware that the glacial epoch was characterized by a sequence of glaciations and deglaciations. At every glaciation water was withdrawn from the ocean and at every deglaciation that same water, or a fraction of it, was restored to the ocean. There was thus a sequence of rises and falls of the sea surface in the latter part of the time in which the Mississippi delta was built. The failure, until recent years, to recognize the magnitude of this alternate rise and fall of sea level in Pleistocene time is an interesting phenomenon of the progress of geological thought. One which should make us blush. A great many geologists, including myself, have seen San Francisco Bay and the Golden Gate. Up till quite recently scarcely a man of them but heaved a great sigh of satisfaction and exclaimed: What a splendid proof of the depression of the coast! And yet, although the bay is certainly flooded from the ocean by way of the Golden Gate, the inference that a depression of the coast caused the flood is not necessary. The rise of the level of the ocean would do the job quite as well. Today it is impossible to

disbelieve that in Recent time the surface of the whole ocean has risen a very notable amount. Two years ago I made a somewhat elaborate computation (1940) of the rise of sea level due to the melting of part of the ice of the last continental glaciation. I came out of the daze with a figure of 426 feet for the measure of the rise. The estimate, of course, lacks precision, and I do not regard it as more than a first approximation. But however it may be discounted there is no escape from the conclusion that the melting of the ice raised the level of the ocean several hundred feet. The deepest sounding in the Golden Gate is 414 feet, and the position of that sounding is probably the bottom of a gorge close to the old shore line. The sea level, when the continental glaciation was at its maximum, must have stood 414 + feet below the present sea level, which is fairly close to 426 feet.

Returning now to the Mississippi delta, the five-fold recurrence of degradation and aggradation of its surface is correlated by Russell with a corresponding recurrence of fall and rise of sea level, and the cause of the latter is ascribed to glaciation and deglaciation. Other workers in the same field, particularly Barton, Frink, and Fisk, are in agreement with him. The stratigraphy is arrived at by the inspection and measurement of formations in natural exposures in the upstream extension of the delta, and following these seaward by means of the record of wells. There are over 100,000 wells, many of them very deep, drilled into or through the delta. The record of these, though not always satisfactory, together with core samples, have in most cases been preserved. A study of these shows that the formations all dip seaward and increase in thickness and in angle of dip in the direction of the dip. The formations referred to are, for the most part, the deposits that were contributed to the delta in times of high sea level, when the process of sedimentation was exceptionally active in what is now the central part of the delta. The increase in thickness of several of these formations, as they are followed down the dip, is very remarkable. Three of them thicken to over 3000 feet each, a fact which impressively proves gradual subsidence *pari passu* with deposition.

Owing to the inconstancy of the stream courses across the surface of the delta—that is, their habits of migrating from one position to another in the course of long periods—the center of maximum deposition also shifts. It results from this that the structure of the delta is not simple, but is made up of an overlapping aggregate of many deltas, each one of which has its own center of accumulation for the time being, and, of course, also its own center of subsidence. But that is a detail upon which I am not qualified to speak. I am concerned here with the embankment as a whole.

But the migration of the streams on the delta surface has much to do with the formation of terraces, and the preservation of these from burial in areas of minimum deposition, as well as in times of high sea level, after the streams have become adjusted to that level. Russell states that the oldest of these terraces slopes seaward at about 10 feet to the mile; the next younger slopes 6 feet, the next 2 feet, and the next 1 foot to the mile. The most recently formed terrace, corresponding to the present high sea level, slopes only half a foot to the mile. These slopes, however, steepen toward the central part of the delta. The slope of the terraces, in excess of about half a foot to the mile, and the increase of that excess with age, is taken to be an expression of the subsidence of the main embankment. But, from what I have already said, such subsidence probably occurs to the seaward of the delta shore, the landward side of that line having arrived at stability of position, except for sinking due to compaction.

The vicissitudes which mark the later stages of growth of the delta of the Mississippi are, as Russell states, referable, with little or no doubt, to the recurrence of glaciation and deglaciation, and the effect of these on sea level, in Pleistocene time. In Tertiary time we have no evidence of glaciation in arctic regions. On the contrary fossil plants in certain parts of the arctic indicate a warm climate. So we may picture to ourselves a preglacial condition of the earth when arctic and antarctic regions were devoid of ice. Then the ocean basins were full to the brim. In the time of maximum Pleistocene glaciation enough water was taken from the ocean to lower its level probably 800 feet below the preglacial position. One effect of this drop of sea level was the emergence of the Mississippi delta and of all other marine deltas outside of the frozen terranes. The surfaces of all these deltas, we may be sure, were deeply scored by stream trenches. This, however, was a temporary condition. The next interglacial epoch was surely recorded in the delta by the converse set of conditions. This alternation of glacial and interglacial climate was recorded in changes of the position of sea level which affected all the coasts of the earth. Thus in the subtropical region of the Gulf of Mexico the relative chronology of the major events of the glacial period are being deciphered far more clearly and definitely than they have ever been in subarctic latitudes.

END OF THE DELTAIC CYCLE

As to the close of the deltaic cycle of the Mississippi there is this to be said: Geologists believe that the most characteristic structural feature of many mountains is the folded condition of sedimentary strata. The folding is in all cases due to horizontal compression, but there are all

degrees of intensity of folding, from broad, shallow undulations to highly compressed, intricately contorted and overturned plications. They have observed also that the thickness of sedimentary strata in the mountains is much greater than elsewhere, and that the dominant feature of many folded ranges is a great syncline, cradled in which are many subordinate synclines and anticlines. The dominant syncline outlines a mountain range and is known as a geosyncline. Such a structure is well exemplified in the Appalachian Mountains, where it has been shown by stratigraphic studies that the thickness of shallow-water beds is much greater in the axial region of the vast troughlike mountain fold than it is on its flanks. It is evident that, in the depositional field, where this body of strata accumulated, the earth's crust subsided to accommodate the deposits. The maximum subsidence, and therefore the maximum volume of sediments, was localized along a zone which became, by mountain-making deformation, the axis of the geosyncline. There can be but little doubt that the subsidence was due to yielding underload, and that, throughout the sinking, crustal balance was maintained by out flowage of heavy rock in depth from beneath the seat of deposition. The maximum thickness of sedimentary strata in the Appalachian Mountains is about 40,000 feet. The attainment of this volume seems to have brought the sedimentary process to an end. The slow subsidence of accommodation to maintain isostatic balance gave place to a more vigorous deformation due to the collapse of the depositional trough under horizontal compressive stress. A mountain-making movement was thus inaugurated. The trough became a geosyncline, and this by continued compression was hoisted high above sea level and sunk deep into the crust. The range thus created was supported by the depression of the keel of the geosyncline. Heavy rock in depth was displaced, and the new range, consisting of light rock, acquired the buoyancy necessary for its existence as a feature of the surface relief. But the range was hoisted into the zone of vigorous erosion and in the course of time became greatly reduced. Its reduction signified a loss and surface redistribution of its mass. The buoyancy of the range then operated for its uplift, so that the lowering of its surface was far less than the mean thickness of the prism removed by erosion. The uplift, of course, also applied to the bottom of the geosyncline, so that the keel is not nearly as deep as it was originally. The key to this sequence of events in the genesis of a mountain range like the Appalachians appears to be that, as accumulation and subsidence proceed, several miles of the upper part of what seismologists call the "granitic" layer are displaced by weak incoherent sediments. The lower part of the "granitic" layer as it sinks approaches the asthenosphere and loses strength as it does so.

The residual upper part may be much less than half the thickness of the normal "granitic" layer, and be so attenuated that it yields to the compressive stress inherent in the crust. The maximum thickness of strata in a single sequence known anywhere appears to be 40,000 to 42,000 feet; and we are justified in the inference that when this thickness is attained further additions are precluded by the inauguration of the mountain-making movement.

The foregoing outline of the genesis of the Appalachian range, and of the antecedent sedimentary cycle, is indicative of the fate that awaits the Mississippi delta. The next mountain range to be developed at the continental margin will have an east-west trend, between south Texas and north Florida. The Mississippi River will have to find another outlet for the drainage of its basin, perhaps between the east end of the new range and the south end of the Appalachians. The present deltaic cycle will have closed. How long it may be in terms of years before the end it is hard to surmise. If the Mississippi delta started in a sea 4 km. deep, it now has a thickness very little less than 40,000 feet. Under that assumption it is evidently approaching the critical stage. But under the same assumption it can never get any thicker, and may be able to withstand the comprehensive stress of the crust until it is much broader than it is now.

SUPPLEMENT

As a supplement to the conclusions arrived at by this study of the Mississippi delta I may add another of general interest. On two oceanographic charts of the earth the depth of the ocean is indicated by submarine contours. The highest of these, at a depth of 200 meters or 656 feet, encloses the continental and insular areas. It includes the extent of off-shore water having less than that depth. The shallow seas thus indicated are particularly noteworthy off the coasts of eastern and southeastern Asia and the East Indies, Australia, Antarctica, eastern North America including the Gulf of Mexico, western India, as well as Bering Sea and the adjoining Arctic Ocean, Hudson Bay, and the Persian Gulf. I have told you that the melting of the ice of the Glacial Period has thus far raised the level of the ocean over 400 feet. But only about half of that ice has been melted. The other half still remains. If it too should melt, by further amelioration of world climate, the surface of the ocean would rise a few hundred feet more. We may safely infer, therefore, that in preglacial time, when there was no ice, the coasts of all the oceans must have been inundated more extensively than they are today. This inundation would of course have determined a configuration of coast line very different from that with which we are now familiar.

The difference would have been most marked in coastal regions of low relief with large river valleys; while bold coasts presenting mountain fronts, like that of the west side of South America, would not have differed much from those of the present time. We may also safely infer that, at the time of maximum glaciation, the greater part of the sea bottom within the — 200 meter contour would have been exposed as dry land, thus adding very considerably to the area of the continents. This addition of flat land to the continental margins would have induced migration of people then living on the earth to the new land surface. They would have moved in slowly as the shore retreated and probably would not have been conscious of the change going on. There they flourished for many centuries, if not millennia, with all the material adjuncts of a primitive social order. By the rise of ocean level, that came with the melting of the continental ice sheets, that migration was reversed. The people moved up the coastal slope in front of the rising tide, also slowly and also probably unconscious of the change that was going on. The material evidences of their existence are now lost to us. Their habitations, their monuments and their graves, together with their art and their refuse, lie scattered over the floor of the ocean. All that is left to us of this flight from the sea is the myth of a universal flood which we have never believed.

REFERENCES CITED

Barton, D. C., Ritz, C. H., and Hickey, Maude (1933) *Gulf Coast geosyncline,* Am. Assoc. Petrol. Geol., Bull., vol. 17, p. 1446.

Dole, R. B., and Stabler, H. (1909) *Denudation,* U. S. Geol. Survey, W. S. Paper 234, p. 78-93.

Frink, J. W. (1941) *Subsurface Pleistocene of Louisiana,* La. Geol. Survey, Bull. 19.

Gilbert, G. K. (1882) *Contributions to the history of Lake Bonneville [Utah],* U. S. Geol. Survey, Ann. Rept. 2, p. 167-200, map.

——— (1890) *Lake Bonneville,* U. S. Geol. Survey, Mon. 1.

Humphreys, A. A., and Abbot, H. L. (1861) *Report upon the physics and hydraulics of the Mississippi River,* U. S. Army, Corps Topogr. Eng., Prof. Papers 4; reprinted 1876, p. 148.

Lawson, A. C. (1934) *The Eparchean peneplain; an exploitation of the doctrine of isostasy,* Geol. Soc. Am., Bull., vol. 45, no. 6, p. 1069-1071.

——— (1938) *Isostasy of large deltas,* Geol. Soc. Am., Bull., vol. 49, p. 401-416.

——— (1940) *Isostatic control of fluctuations of sea level,* Science, August 23, p. 162.

Russell, R. J. (1940) *Quaternary history of Louisiana,* Geol. Soc. Am., Bull., vol. 51 p. 1199.

University of California, Berkeley, Calif.
Manuscript received by the Secretary of the Society, June 8, 1942.